1970

N. E. I. G. C.

RANGELEY LAKES — DEAD RIVER BASIN REGION

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

GUIDEBOOK

for field trips in

The Rangeley Lakes - Dead River Basin Region,

Western Maine

Gary M. Boone

Editor

62nd Annual Meeting

October 2, 3, and 4, 1970

University of New Hampshire

New England Intercollegiate Geological Conference (N.E.I.G.C.)

The N.E.I.G.C. was begun in 1901 as an informal field trip, organized by William Morris Davis, to the Connecticut Valley of Western Massachusetts. The 1970 Conference in Rangeley marks the 62nd annual meeting (and 6th in the State of Maine). Throughout its history the sole purpose of the N.E.I.G.C. has been to bring together in the field those geologists interested and active in New England Geology, to consider and discuss the results of new mapping and other geologic studies.

D.W. Caldwell, Secretary, N.E.I.G.C.

Conference Organization

Gary M. Boone
Eugene L. Boudette — Co-chairmen
Robert H. Moench

Field trip leaders and guidebook authors:

G.M. Boone, Department of Geology, Syracuse University, Syracuse, N.Y.
H.W. Borns, Department of Geological Sciences, University of Maine, Orono, Maine
E.L. Boudette, U.S. Geological Survey & Dartmouth College, Hanover, N.H.
P.E. Calkin, Department of Geology, State University of New York at Buffalo, N.Y.
R.G. Doyle, State Geologist, Maine Geological Survey, Augusta, Maine
R. Fournier, Department of Geology, University of Cincinnati, Cincinnati, Ohio
J.C. Green, Department of Geology, University of Minnesota, Duluth, Minnesota
C.V. Guidotti, Department of Geology & Geophysics, University of Wisconsin, Madison, Wisconsin
D.S. Harwood, U.S. Geological Survey, Washington, D.C.
C. Koteff, U.S. Geological Survey, Boston, Massachusetts
R.H. Moench, U.S. Geological Survey, Denver, Colorado
F. Pessl, Jr., U.S. Geological Survey, Boston, Massachusetts
J. Raabe, Northwestern Connecticut Community College, Winsted, Connecticut
W.W. Shilts, Geological Survey of Canada, Ottawa, Canada

Price of guidebook: $4.00 (U.S.)
Requests for orders may be addressed to:

Gary M. Boone
Department of Geology
Syracuse University
Syracuse, New York 13210

TABLE OF CONTENTS

PAGE

Foreward and Acknowledgements

This fall we converge in Maine in a year that coincides with the State's sesquicentennial. The present conference, offered in the spirit of a wilderness experiment, is being held in surroundings that are still relatively rustic, and where several individuals, rather than an institution, collectively act as host.

Those in the vanguard of modern geological mapping and topical research here began their work in 1956. The results of their work, like ours, are only partially published or are still in progress.

One hundred ninety-five years ago, another group, just as determined, and perhaps as motley, came through the Dead River region on another wilderness experiment. They trekked northwestward, were beset by a hurricane, used bateaux where they could, and in that earlier winter of discontent, history reminds us that they had some difficulty when they reached Québec City. For our present Conference, however, we are concerned with the history of southeastward-encroaching ice-sheets and of deformed rocks in which much of the structure, though not all, gives southeast-facing successions of layered rocks. You will not be traversing continuously southeastward on the various trips, but the traditional purpose of the New England Intercollegiate Geological Conference, which is to spur the discussion of field geologic relationships, ensures by its very nature a more constructive approach to progress in mutual understanding than the objectives of the northwestward-marching group of nearly two centuries ago.

Since the times of Benedict Arnold and Timothy Bigelow, and of William King, Maine's first governor, the region in which the Conference convenes has become widely known for its woods products-oriented industries, and for its sporting spirit in many respects. In the beginning of this 150th year since Maine severed the cord with its Commonwealth neighbor to the south, the questionable advance of civilization has done little to erode the elemental solidarity of the Downeast viewpoint.[1]

The region is one of special bedrock stratigraphic significance, for it forms a bridge between geology which is interpreted with relative confidence in Maine, Québec, and the Maritime Provinces to the northeast, and in New Hampshire and Vermont to the southwest. To the northeast the metamorphic grade decreases, and primary depositional features and fossils are better dispersed. To the southwest regional grade increases and discoveries of critical fossil localities have been fewer. This generality would not be news to the 19th Century geologists Hitchcock or Jackson, but the problems and potential answers lurking in the metamorphic transition are still key factors with us today.

Growing interest in the Pleistocene history of the region is evident in studies that are extending the more clearly documented histories in the Atlantic and St. Lawrence lowlands headward toward the Boundary Mountains watershed. It may be that a C^{14} 'spike' will soon

---

1. Cf. Strunk, Jud, 1969, Downeast Viewpoint: Columbia, CS9990; 2 sides.

be driven to unite the approaching paths of research on each side of the international boundary.

To acquaint students and professional geologists with a regional geologic setting, introductions to the bedrock and surficial geology precede the separately authored discussions of field trip localities.

Because the Rangeley Lakes-Dead River Basin Region is rural, and much of it forested, logistic responsibility for the conference has been divided among the hosting group of leaders to cope with the various demands and constraints of tall-timber geology. Although some inconvenience may be experienced in transportation, shelter, and quest for sustenance, the hosts believe that this will be more than repaid in terms of the region's October scenery, pure air and water, spectacular outcrops, and friendly residents.

None of the conference organizers nor, with one exception, scarcely any two trip leaders reside in the same part of the country. Nevertheless, though instant communication has oft-times been difficult, the ready cooperation of all colleagues in this venture is gratefully appreciated.

The work of Eugene Boudette and Robert Moench, as co-organizers, has been a key factor in accomplishing the many and varied tasks. To them, and to Tina Cotton and others of the U.S. Geological Survey offices in Boston, I offer grateful thanks. The generous cooperation of Lincoln R. Page in these respects is very much appreciated. I also acknowledge the support of the Maine Geological Survey toward the operation of the Conference. The departments of geology of Dartmouth College and of Syracuse University rendered helpful assistance. Susan Stores, of the Department of Geology, Syracuse, helped greatly in the organization of manuscripts and typing much of the guidebook.

We are mindful that the Town of Rangeley has done much to accommodate our needs, as have other residents and land-owners of the region. Their gracious support has been vital to the Conference.*

Gary M. Boone

Editor

* I am very grateful to the Maine Sesquicentennial Commission for their recent donation in support of the publication of this guidebook, and for their permission to use the Sesquicentennial seal.

Bedrock geology of the Rangeley Lakes-Dead River basin region, western Maine

by

Gary M. Boone[1], Eugene L. Boudette, and Robert H. Moench[2]

Introduction

Northwestern Maine forms a transitional zone through which the fundamental structure and stratigraphy of the less metamorphosed rocks to the northeast in Maine and Canada can be correlated with the highly metamorphosed, structurally complex rocks of southern New England. In this region (fig. 1) a nearly complete stratigraphic section is exposed from Precambrian(?) through the Lower Devonian; intrusive rocks range in age from Early Ordovician(?) to Triassic(?).

In this introduction we hope to present a synthesis of the stratigraphy and structure of the area; more detailed accounts are presented in the individual field trips. Expanding lumber operations within the past decade have opened up previously inaccessible areas, but several key exposures still remain beyond reasonable reach for conference-sized groups.

The area is large and geologically complex, and sufficient diversity of interpretation exists to provide the dynamism appropriate for a conference of this sort. Where points of contention occur in the interpretation, we hope we have presented the respective cases fairly. Modern detailed geologic investigation of the region began in 1948 when Arthur J. Boucot, then a graduate student at Harvard University and later a member of the U.S. Geological Survey, undertook his classic study of the stratigraphy of the Moose River synclinorium (Boucot, 1961, 1969; Boucot and others, 1964). C. Wroe Wolfe of Boston University began a long-term study of the southeastern part of the region in 1948, which was carried on more recently by Mohammed A. Gheith, also of Boston University. They directed an undergraduate field camp, and N.S.F.-N.A.G.T. summer institute, and graduate students in several mapping projects: Robert H. Moench (Phillips quadrangle), Robert J. Willard (Kennebago quadrangle), Victor Columbini (Rangeley quadrangle), Ross G. Schaff (Little Bigelow Mountain quadrangle), and Stanley Skapinsky (Kingfield quadrangle). Although work of the Boston University group is largely unpublished, it has stimulated further research and contributed greatly to our present geologic knowledge of the region.

---

[1]Maine Geological Survey and Syracuse University, Syracuse, New York

[2]U.S. Geological Survey, Boston, Massachusetts, and Denver, Colorado

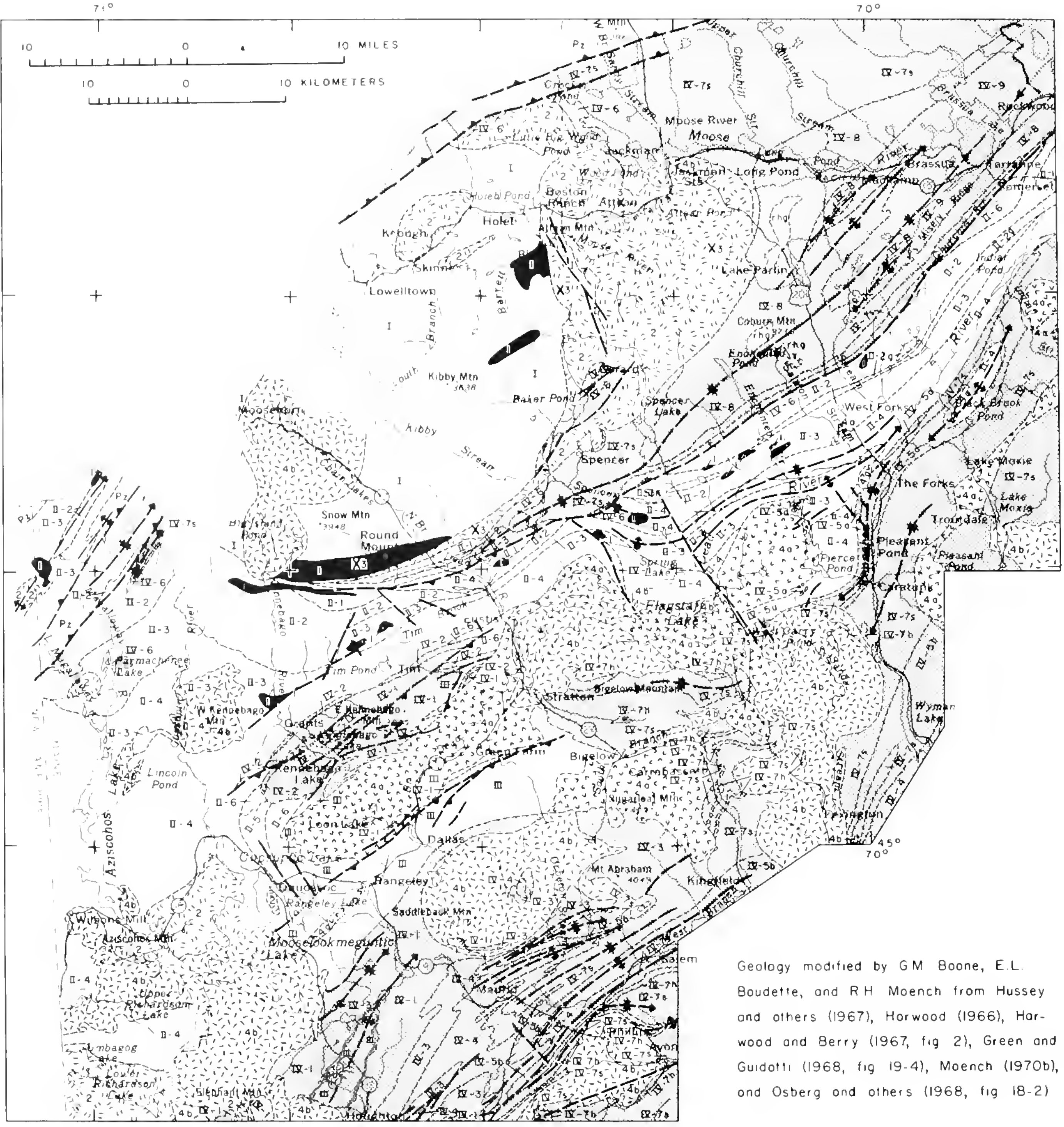


Figure 1. Bedrock geologic map of the Rangeley Lakes-Dead River basin region, western Maine

EXPLANATION

Rocks of sedimentary and extrusive origin metamorphosed in much of the region to chlorite grade and higher grades of metamorphism

Merrimack Synclinorium — Boundary Mountains Anticlinorium — Connecticut Valley-Gaspé Synclinorium

Lower to Middle Devonian

Lower Devonian and Lower Devonian(?)

Moose River Synclinorium

IV-9

Tomhegan Formation
Metasandstone, metapelite, and felsic to intermediate volcanic rocks

IV-8

Tarratine Formation
Metasandstone and metapelite with subordinate quartzite and metalimestone

IV-7h / IV-7s

IV-7s: Seboomook Formation and probable equivalents in southeastern and northeastern strike belts. Cyclically bedded metashale and metasandstone

IV-7h: Hildreths Formation and equivalent calc-silicate and metasandstone units; may be IV-5 in part, east of Stratton

IV-5b / IV-5a

IV-5a: Calcareous metapelite and calc-silicate rock, Pierce Pond and northeastward

IV-5b: Madrid Formation and unnamed calcareous metasandstone and metapelite near Wyman Lake, Bingham quadrangle

Lower Silurian

IV-6

Formations of the Moose River synclinorium and unnamed units near Jackman and to the southwest
Calcareous slate and metalimestone

IV-4

Smalls Falls Formation and unnamed units near Wyman Lake, Bingham quadrangle
Rusty-weathering, sulfidic metashale, metasandstone, and metaconglomerate

Pz

Light-hued phyllite and slate, metasandstone, metagraywacke, pillow lava, and metavolcanic rock

Lower Silurian and Lower Silurian(?)

IV-3

Perry Mountain Formation
Bedded metasandstone, metashale, and metaconglomerate

IV-1

Rangeley Formation
Metashale, metasandstone, and metapelite
Unnamed units near Spring Lake in the Stratton quadrangle
Banded calc-silicate rock

IV-2

Vitreous quartzite, locally conglomeratic, feldspathic metasandstone, and metapelite

?

TACONIC UNCONFORMITY

Upper Ordovician(?)

III

Quimby and Greenvale Cove Formations
Metagraywacke, felsic metavolcanic rock, metapelite, and metasandstone

Middle Ordovician

II-6 / II-5

II-6: *Dark sulfidic slate, metagraywacke, and local metaconglomerate*

II-5: *Greenstone*

Upper Cambrian(?) to Middle Ordovician

II-4

Green, locally variegated red and green metapelite and subordinate green metagraywacke; *characteristically pinstriped; laminated to thickly bedded*

II-3

Dark sulfidic slate and phyllite, metagraywacke, and feldspathic metasandstone

II-2g / II-2

II-2: *Green, quartzose, chloritic metagraywacke and interlayered metapelite, red slate, and greenstone*

II-2g: *Greenstone of II-2 shown differentiated east of Spencer Lake*

II-1

Dark and light greenstone, felsic metavolcanic rock, and cherty iron formation

UNCONFORMITY

1

Chain Lakes massif
Granofels, gneiss, schist, quartzite, and thinly layered amphibolite

Intrusive Rocks

Triassic(?) lamprophyre dikes, widespread throughout the region, are not shown

Devonian and Devonian(?)

4b

Quartz monzonite and granodiorite

4a

Norite, troctolite, gabbro, diorite, and quartz diorite

rhg

Garnet rhyolite and felsic hypabyssal rocks in the Moose River synclinorium

Silurian and Silurian(?)

Ordovician and Ordovician(?)

x3

Quartz porphyry and aplitic rocks

2

Altered quartz monzonite and granodiorite; includes Attean Quartz Monzonite

Cambrian or Ordovician

1

Altered epidiorite, serpentinite, quartz diorite, pyroxenite, gabbro, and alaskite of the ultramafic complex

Precambrian(?)

Contact
Dotted where concealed

Unclassified

Thrust or reverse
Sawteeth on upper plate

Faults

Syncline

Anticline

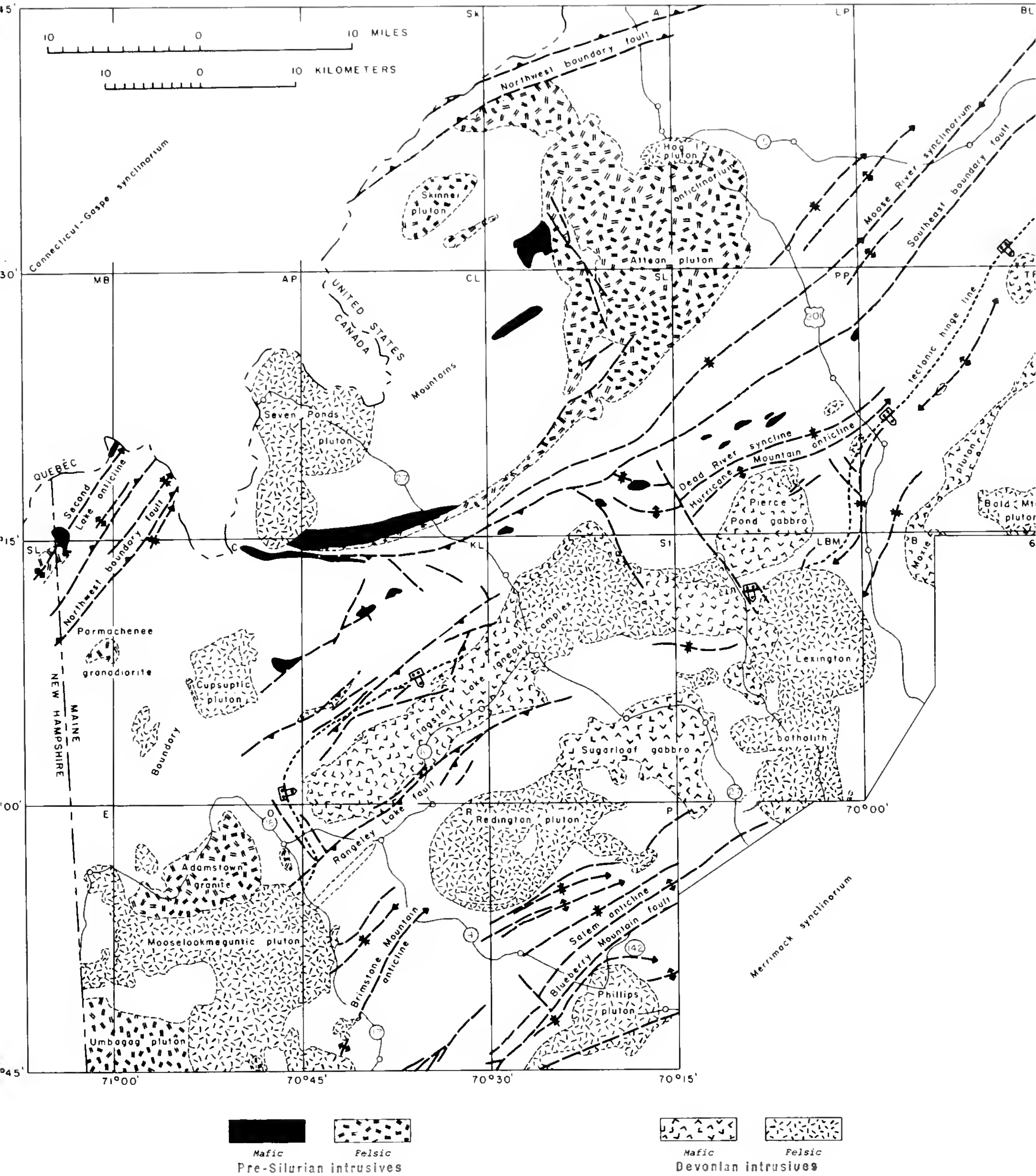
Folds
Direction of plunge shown where known

Patrick M. Hurley and James B. Thompson, Jr. (1950) conducted an aeromagnetic-geologic reconnaissance over the northeastern part of the region.

In 1956, the U.S. Geological Survey began a broadly based research program oriented toward the economic geology of the region. This program is still in progress. It includes detailed geologic mapping at the scale of 1:62,500, geochemical and geophysical exploration, aeromagnetic mapping, and heavy-metals resource studies. Geologic quadrangle mapping that has been completed or is in progress by both Survey and non-Survey geologists is summarized in figure 2. Frank C. Canney has carried out research in geochemical exploration and remote-sensing techniques since 1958 when Edward V. Post and others were conducting geochemical reconnaissance studies. Martin F. Kane and others carried out regional gravity field studies between 1959 and 1969. Aeromagnetic maps for much of the region have been published by the U.S. Geological Survey. (Bromery and others, 1957a, 1957b; Bromery, Soday and others, 1963; Bromery, Tyson, and others, 1963; Bromery and Gilbert, 1962; Henderson and others, 1963a, b; Henderson and Smith, 1964; Boynton and Gilbert, 1964a, b, c, d; U.S. Geological Survey, 1969).

Fossils in metamorphic rocks of the area have provided a key to fit the regional stratigraphy into the geologic time scale. We are especially indebted to the following paleontologists for working with the partially metamorphosed fossils that we and others have collected: Arthur J. Boucot, Oregon State University: William B.N. Berry, University of California at Berkeley; William H. Oliver and Robert B. Neuman, U.S. Geological Survey; and Robert Finks, University of Miami in Florida. We are also indebted to many others, only a few of whom are mentioned below, for their beneficial contributions through discussion and encouragement. Lincoln R. Page of the U.S. Geological Survey has been a prime moving force behind modern geologic mapping and correlation here and elsewhere in New England. A.J. Boucot established the Silurian and Devonian stratigraphy of the Moose River synclinorium as we now know it. Raymond A. Marleau (1968) mapped the Woburn, East Megantic, and Armstrong areas, Quebec, to the northwest while a graduate student at Laval University. Marland P. Billings and J.B. Thompson, Jr., of Harvard University, John B. Lyons of Dartmouth College, Arthur M. Hussey, II, of Bowdoin College, and Robert G. Doyle, State Geologist of Maine, have made significant contributions to the regional stratigraphy through their involvement with similar problems of New England geology. Finally, we are grateful for the interest and cooperation extended by the principal landholders in the region, including the Brown Company, Scott Paper Company, Hudson Paper Company, Central Maine Power Company and affiliates, and the Dead River Company.

The geologic nomenclature used in this report is from many sources and does not necessarily reflect the usage of the U.S. Geological Survey.



Sk, Skinner - *E.L. Boudette (unpublished data)*
A, Attean - *A.L. Albee (1961), A.L. Albee and E.L. Boudette (1971)*
LP, Long Pond - *A.J. Boucot (1961, 1969)*
BL, Brassua Lake - *A.J. Boucot (1961, 1969)*
MB, Moose Bog - *J.C. Green (1968)*
AP, Arnold Pond - *D.S. Harwood (1968)*
CL, Chain Lakes - *E.L. Boudette (unpublished map)*
SL, Spencer Lake - *A.J. Boucot (1961, 1969), E.L. Boudette and G.M. Boone (unpublished data)*
PP, Pierce Pond - *A.J. Boucot (1961, 1969), G.M. Boone (unpublished map)*
TF, The Forks - *E.V. Post (unpublished map)*

SL, Second Lake - *J.C. Green (1968)*
C, Cupsuptic - *D.S. Harwood (1966)*
KL, Kennebago Lake - *E.L. Boudette (unpublished map)*
St, Stratton - *Andrew Griscom (unpublished map)*
LBM, Little Bigelow Mountain - *G.M. Boone (1970)*
B, Bingham - *B.D. Keith (unpublished map), G.M. Boone (unpublished map)*
E, Errol - *J.C. Green (1964)*
O, Oquossoc - *C.V. Guidotti (unpublished map)*
R, Rangeley - *R.H. Moench (1970)*
P, Phillips - *R.H. Moench (1970)*
K, Kingfield - *John Raabe (unpublished map)*

Figure 2. Index map of major tectonic features showing locations of 15-minute quadrangles and sources of detailed and reconnaissance bedrock geologic mapping in the Rangeley Lakes-Dead River basin region, western Maine.

Structure

Major tectonic features

Thirteen major structural elements dominate the regional east-northeast tectonic grain (fig. 2). From northwest across strike to the southeast these structures include the:

Connecticut Valley-Gaspé synclinorium
Second Lake anticline
Northwest boundary fault
Boundary Mountains anticlinorium
Moose River synclinorium
Southeast boundary fault
Dead River syncline
Hurricane Mountain anticline
Rangeley Lake fault
Brimstone Mountain anticline
Salem anticline
Blueberry Mountain fault
Merrimack synclinorium

The Boundary Mountains anticlinorium (Albee, 1961; Green and Guidotti, 1968) forms the central structure which is complemented on the northwest by the Connecticut Valley-Gaspé synclinorium (Cady, 1960) and on the southeast by the Moose River synclinorium (Boucot, 1961) and the Merrimack synclinorium (Billings, 1956; Osberg and others, 1968). The Moose River and Merrimack synclinoria are separated by Ordovician(?) low-rank metamorphic rocks and by the southeast boundary fault that truncates the southeast limb of the Moose River synclinorium. The structure of the Ordovician rocks is more homoclinal to the southwest and more anticlinorial to the northeast. Both the northwest and southeast boundary faults are primarily thrust or reverse faults with an appreciable strike-slip component. Numerous subsidiary folds and faults, local unconformities, short-range facies changes, and major wedge-shaped depositional prisms complicate the fundamental structure. The central part of the region contains a northeast-trending tectonic transition zone (Pavlides and others, 1968) called a hinge line. The Taconic unconformity is recognized only northwest of the hinge line. Southeast of the hinge line, Moench (1969) found more or less uninterrupted sedimentation between Middle Ordovician and Silurian rocks.

Major stratigraphic divisions

The four major stratigraphic divisions recognized in the area are based on criteria such as unconformable relations of metamorphic grade, degree of internal deformation, lithologic assemblages, and paleontology. These divisions are:

(I) Precambrian(?) crystalline rocks (Chain Lakes massif) of originally pervasive high metamorphic rank.

(II) Upper Cambrian(?) to Middle Ordovician eugeosynclinal assemblage generally ranging from chlorite to staurolite grades of regional metamorphism.

(III) Late Ordovician(?) eugeosynclinal assemblage with metamorphic grade similar to II.

(IV) Lower Silurian to Lower Devonian miogeosynclinal and eugeosynclinal assemblages which are chlorite to sillimanite and K feldspar metamorphic grade.

Stratified rocks in the chlorite zone of regional metamorphism (east-central part of the region, fig. 1) are locally involved in higher grades of dynamothermal, low-pressure facies progressions (Childs and Boone, 1968; Green, 1963) typical of the Buchan of Scotland, or Bosost area of the Pyrenees (cf. Winkler, 1967, p. 125-130). To the south and southwest, strata were dynamothermally metamorphosed at higher pressure levels typical of the Barrovian, or Dalradian series (Evans and Guidotti, 1966; Moench, 1970).

Rocks of varied chemistry and emplacement mechanics assigned to three and possibly four plutonic episodes intruded the metamorphic rocks in: Late Cambrian(?) to Middle Ordovician, Late Ordovician(?), Early to Middle Devonian, and Triassic(?) times.

The axial area of the Boundary Mountains anticlinorium is underlain mainly by stratigraphic units of divisions I and II with local unconformable outliers of IV. The recognition of the Chain Lakes massif (I) in the anticlinorium as separate from and older than Cambro-Ordovician rocks (II) presents an interpretation different from that recently made by Green and Guidotti (1968), Harwood (1968), and Albee (1961) who correlated I and II. Our interpretation however, agrees with Boucot (1961, 1969) and Boucot and others (1964). Field Trip C deals specifically with this topic.

The opposing flanks of the Boundary Mountains anticlinorium do not expose symmetrical stratigraphic sequences or comparable subsidiary structures. Lower Devonian rocks of division IV correlate across the crest of the anticlinorium, and units of division II have been mapped around the southwest-plunging crest of the anticlinorium and projected on regional strike along the northwestern flank (Harwood and Berry, 1967; Harwood, 1969). A major unconformity beneath rocks dated Silurian or younger defines the core of the anticlinorium (Albee, 1961). This unconformity, which documents the Taconic break has not been recognized in the Merrimack synclinorium or in the Connecticut Valley-Gaspé synclinorium. An unconformity older than the Taconic separates rocks of the Chain Lakes massif from those of division II. Faulting and intrusive rocks of Devonian age obscure this older unconformity to the northwest. The southeast flank of the anticlinorium has more varied distinctive lithologies and is presently known in more detail. Lithologic units form through-going strike belts or are draped across the nose of the anticlinorium. Despite varied tectonic events and widespread polymetamorphism, unique combinations of depositional structures and lithic sequences persist within units.

Merrimack synclinorium

The stratified rocks facing the axial trace of the Merrimack synclinorium southeast of the Chain Lakes massif are assigned to an older depositional prism (divisions II and III), and to a younger prism (division IV) which conformably overlies the older prism where the Taconic break is absent. In ascending order, from northwest to southeast, the older prism contains: greenstone; metamorphosed quartz-latitic volcanic rocks, metagraywacke, metapelite, and metamorphosed cherty iron formation; euxinic metapelite and metasandstone; chloritic metapelite and calcareous quartzite; and euxinic metapelite containing Middle Ordovician fossils, greenstone, metagraywacke, felsic metavolcanic rocks; euxinic metasandstone, metapelite, metaconglomerate, felsic metavolcanic rocks, and laminated metasiltstone. This nearly homoclinal structure passes northeastward into the Dead River syncline and the Hurricane Mountain anticline. The northwestern boundary of the younger prism is marked by the presence of orthoquartzite, metasandstone, and slate (IV-2) of late Early Silurian age which is correlated with the Clough Formation of New Hampshire (Billings, 1956 and Boucot and Thompson, 1963). The slates and younger metapelites are characteristically less magnesian than those of divisions II and III. This compositional boundary is essentially coincident with the southeasternmost extent of the Taconic unconformity, where the uppermost unit of the older prism becomes conformable with rocks of division IV. The stratigraphic sequence in the younger prism is in ascending order metapelite, metasandstone, and metaconglomerate of the Rangeley Formation (Moench, 1970; Osberg and others, 1968); cyclically bedded metasandstone and metapelite; euxinic metapelite and metasandstone; calcareous metasandstone; and metapelite of the Seboomook Formation.

The stratigraphic sequence of both prisms is repeated to varying degrees by a combination of folds and faults of regional scale including the southeastern boundary fault which repeats nearly all the section of division II south of Round Mountain. The structural interpretation presented here is largely based upon detailed tracing of lithofacies units and observation of top-facing directions, especially at contacts. Where fault repetition of division II rocks occurs on the southeastern boundary fault, the total section in II is observed to thicken remarkably toward the southeast. The same observation is true of the units in divisions III and IV, many of which thin and pinch out northwestward over the tectonic hinge line. Our conclusion is that the geosynclinal furrow receiving the sedimentary and volcanic protoliths of these rocks existed at least as early as Cambrian time.

Connecticut Valley-Gaspé synclinorium

Recent recognition of widespread units of division II in the Second Lake anticline, and fault dislocations nearby (Harwood, 1969) require a reappraisal of previous interpretations (Albee and Boudette, in press; Marleau, 1968; Green, 1968; Green and Guidotti, 1968; and Albee, 1961). The rocks here are not as clearly grouped into depositional prisms, and distinctive basal units are absent. The Taconic unconformity is exposed north of Aziscohos Lake but becomes obscured by the northwest boundary fault to the northeast. Rocks mapped directly above the unconformity are probably

Early Silurian(?) in age and are younger than their counterparts on the southeast flank of the Boundary Mountains anticlinorium. The extent of the unconformity toward the axis of the Connecticut Valley-Gaspé synclinorium is not established. If a relationship comparable to that of the Merrimack synclinorium exists, then rocks (Pz) older than the Early Silurian(?) rocks could be present, although no obvious lithofacies correlations are presently evident. The Compton Formation (Marleau, 1968), a correlative of the Seboomook Formation, is known to thicken to the northwest, and, therefore, sedimentary prisms may also apply to the northwest flank. Time-coincident environments of deposition probably vary on opposite flanks of the Boundary Mountains anticlinorium, and, until more data are available, it can only be observed that rocks younger than those of division II have less lithologic variation and include a widespread abundance of gray or green metapelite and metasandstone. Rocks of division II are correspondingly areally restricted, and the existence of division III rocks is, for the present, an equivocal matter in the Connecticut Valley-Gaspé synclinorium. Harwood and Berry (1967) correlated an euxinic metapelite on the northwest flank of the Boundary Mountains anticlinorium with a paleontologically dated, lithologically comparable unit on the southeast. This correlation is not entirely consistent with observations on the southeast in division II rocks, but the possibility remains that some euxinic units are indeed equivalent and occur in structural windows.

If an older prism is present on the northwest flank of the Boundary Mountains anticlinorium, it may be represented by greenstone and feldspathic metagraywacke in an upright sequence facing the Connecticut Valley-Gaspé synclinorium. The upright sequence of a possibly younger prism may include green metapelite and calcareous metasiltstone, calcareous metapelite and marble, metamorphosed quartz-latite volcanic rocks, red and green metapelite, and cyclically bedded gray slate of division IV. The layered rocks on the northwest are also tectonically repeated, and because of faulting, segments of the section are probably not exposed.

Internal deformation, paleotectonics, and metamorphism

The internal deformation of rock units within the major divisions is variable. This is especially true for the rocks of IV. Crowding adjacent to postkinematic Devonian plutons has produced very complex deformation patterns (see Harwood, 1966; Boone, in press), but other local, complex deformation is not clearly attributable to the emplacement of young plutons, and it must be concluded that either multiple orogeny or a combination of tectonic events, such as major submarine sliding within accumulating sedimentary prisms, may have occured (Moench, 1966, 1970).

Magnesium-rich metapelite and metagraywacke of II, which have been correlated with the Albee Formation by Harwood and Berry (1967) and Green and Guidotti (1968), consistently show intricate folding. This subject is dealt with in detail in Trip H. This may be attributed to refolding of Taconian folds by Acadian tectonism (Harwood, 1966). Thick-layered metavolcanic rocks and massively-bedded graywacke, which are conformable with

and stratigraphically beneath this intensely deformed unit, show more open folding and tilting of fault blocks, probably because of bulk competency that led to buttressing mechanics.

Rocks of IV in the Merrimack and Connecticut Valley-Gaspé synclinoria are tightly folded and faulted along northeast-trending axes. In contrast, units of this group on and adjacent to the anticlinorial tract are much less deformed by folding but locally show late fault breakage and imbrication. Pre-Devonian intrusive rocks are abundantly altered and deformed, whereas the Devonian intrusive rocks are only locally faulted and deuterically altered along their margins. Polymictic conglomerates in IV contain clasts of resistant varieties of rocks and of the distinctive blue quartz of recognizable provenance in divisions II and III. Clasts from division I were not observed in the conglomerates; units of division I were probably not as emergent in Early Silurian time as they were in the Late Silurian when units of IV were deposited directly upon I. The postulated older unconformity separating the rocks of I and II was initially modified by the extrusion of mafic volcanic rocks directly upon granofels of the Chain Lakes massif. This relationship, shown by rocks near Round Mountain, was not as spectacular as the Taconic unconformity, which is commonly marked by distinctive clastic units or by well-bedded calcarenite. Initial contrasts that may have existed beyond the present contrast in metamorphic grade have been partially obscured by the conformable emplacement of the ultramafic complex along the break. At present, only one exposure is known where the direct metamorphic contrast can be seen (see Trip C).

Steeply dipping axial-plane cleavage characterizes all rocks of divisions II, III, and IV. Rocks of I have no axial-plane cleavage, but bedding foliation is common in the well-layered units. The cleavage in the younger divisions is locally folded or cut by slip cleavage. Slip cleavage and fold bands of diverse orientation are locally developed near faults and plutons.

The tectonic and metamorphic overprint of the Acadian orogeny dominates the region by the large scale rearrangement and metamorphism of the rocks. Whether rocks of division II were prograded wholly or in part in Acadian time is not certain. The tectonic environment in Early Ordovician time which accommodated the emplacement of the ultramafic complex and the Attean Quartz Monzonite suggests at least a local metamorphic event that may have resembled that in modern blueschist terranes such as in California. Devonian plutons have been variously dated by radiometric methods as 360 to 395 m.y. (Lyons and Faul, 1968; Faul and others, 1963). Well-defined contact aureoles without major retrograde effects surround many of the plutons that clearly truncate Acadian folds. These plutons were emplaced, therefore, after the principal Acadian deformation and at or slightly later than the thermal peak of regional metamorphism.

Stratigraphy

We use established stratigraphic names. Where stratigraphic problems may exist in lithologic description, sequence, or age from the type area, we briefly mention their history. The stratigraphy is reviewed within the tectonic framework discussed above, although tectonic and stratigraphic boundaries do not always coincide.

Core of the Boundary Mountains anticlinorium

Chain Lakes massif (division I). This unit is composed of the following rock types, any two or more of which are commonly interlayered in a given outcrop: Light-colored, massively bedded quartzite; black schist and massive feldspathic metasandstone interlayered with dark thinly layered amphibolite; gray biotite gneiss; gray well-layered or complexly contorted chloritic quartz-feldspar-muscovite (± sillimanite) schist and felsic metavolcanic rocks; gray massive chloritic quartzfeldspar-muscovite (± biotite ± sillimanite ± garnet ± cordierite) gneissic granofels containing lithic fragments and quartz nodules; and massive dark mafic breccia. The age of the massif has not been established. The unit is probably more than 10,000 feet thick.

Considerable debate surrounds the stratigraphic interpretation of the unit. Harwood (1969) and Green and Guidotti (1968) have recently correlated it with the Dixville Formation of Green (1964) of Ordovician age, preserving an earlier interpretation (Albee, 1961) that the unit is represented in New Hampshire by the Ammonoosuc and Partridge Formations. We believe that relative age of units along the southeastern boundary of the massif in the Chain Lakes quadrangle do not permit this correlation and that the massif lacks any clearly recognized regional counterpart. The massif includes part of the Arnold River Formation of Marleau (1968), which Marleau correlated with the Quebec Group of Cambro-Ordovician age. L.R. Page (personal commun., 1969) believes that the massif bears lithologic similarities to Cambrian(?) units in southeastern New England. We have provisionally assigned the unit an age of Precambrian(?) (see also Boucot, 1961, 1969; Boucot and others, 1964).

Upper Cambrian(?) to Middle Ordovician eugeosynclinal rocks (lower part, division II). This group includes the following lithic types in ascending stratigraphic order: (II-1) Largely massive greenstone and amphibolite with relict pillow structure and breccia preserved locally; the mafic rocks are interlayered with metamorphosed quartz latite flows and breccia near Jim Pond in the Chain Lakes quadrangle. The unit contains lenses of jasper and cherty iron-formation. (II-2) Light-green massively bedded quartz-chlorite-rich, metagraywacke, locally conglomeratic, grading to chloritic phyllite and slate, and, locally containing lenses of brick-red phyllite and slate, and greenstone. (II-3) Dark sulfidic laminated phyllite and slate interbedded with dark metagraywacke and sparse, but massively bedded light-colored feldspathic metasandstone and quartzite. This part of division II ranges in thickness from 8,000 to 15,000 feet.

Rocks of this group include, in part, the Dixville Formation of Green (1964), which has been correlated with the Partridge and Ammonoosuc Formations (see above) and assigned a Middle Ordovician age by Harwood and Berry (1967) on the basis of a dated, lithologically similar (but not contiguous) unit near Oquossoc, Maine. It is important to point out here that a stratigraphic order inversely related to ours has been interpreted by Green (1964, 1968), Green and Guidotti (1968), Harwood (1966, 1969), and Harwood and Berry (1967) which preserves the sequence of the proposed New Hampshire equivalents (Billings, 1956). We feel this interpretation is not consistent with stratigraphic top-facing directions observed at contacts in the central and eastern part of the region and furthermore, that the assigned Middle Ordovician age is not compatible with radiometric ages of intrusive rocks of Early Ordovician or older age (Holmes' scale) which cut the group. Long-range regional correlations with similar rocks in Quebec (St. Julien, 1967) which are cut by similar pre-Silurian intrusive rocks, the youngest of which is dated 480 m.y. (see Poole and others, 1963), indicates that a Cambrian or Early Ordovician age is more likely. We feel this latter interpretation of age is the more reasonable age rather than the currently accepted Middle Ordovician age.

Upper Cambrian(?) to Middle Ordovician eugeosynclinal rocks (upper part, division II). Green, red, locally green and red variegated, and black chloritic phyllite and slate interbedded with subordinate amounts of calcareous metagraywacke and locally massive calcareous quartzite, felsic metavolcanic rocks, and small greenstone lenses occur in the western part of unit II-4. The rocks are laminated to thickly bedded and characteristically pinstriped parallel to the foliation and relict bedding. White laminae range from quartz rich to plagioclase rich across a wide spectrum of metamorphic rank. Well-preserved sedimentary structures in the eastern part of the region indicate a history of redeposition of clastic sediment, reflecting sources characterized in part by mafic volcanic rocks (Trip B-1). The estimated thickness of unit II-4 ranges from 5,000 to 15,000 feet.

This sequence has been assigned in part to the Albee Formation and in part to the Aziscohos Formation of Green (1964, 1968). Harwood (1966, 1968) and Harwood and Berry (1967) assigned an age of Middle Ordovician(?) or older, and interpreted it to be the oldest stratigraphic unit in the region (see also Albee, 1961; Green and Guidotti, 1968). Although the latter interpretation preserves the essential aspects of the New Hampshire pre-Silurian section (Billings, 1956), it is not in accord with the observed sequential relationships east of the Cupsuptic quadrangle. Depositional structures in the gradational contact zone between units II-3 and II-4 in the eastern part of the map area indicate that II-4 is younger than II-3. Furthermore, these rocks, which form a continuous strike belt with the Dixville and Albee Formations, do not represent the oldest pre-Silurian units in the eastern part of the map area.

Upper Silurian(?) miogeosynclinal rocks (middle part, division IV). Rocks of IV-6 occur as scattered outliers of polymictic conglomerate, quartz pebble conglomerate and quartzite, limestone conglomerate, argillaceous or arenaceous limestone and calcareous sandstone, quartzite, and pelitic rocks with sparse felsic volcanic rocks in synclinal keels or block-

faulted structures. Thicknesses as much as a few hundred feet are recorded. All these rocks are fossiliferous with locally abundant shell-fauna assemblages which date from late Llandovery to Ludlow.

Northwest limb of the Merrimack synclinorium

Middle Ordovician eugeosynclinal rocks (upper part, division II). The rocks exposed primarily north of Rangeley Lake include in approximate ascending order massive to poorly foliated greenstone (II-5) and euxinic sulfidic black slate with abundant thin beds of metagraywacke (II-6). This sequence may be as much as 13,500 feet thick. Graptolites indicative of Zone 12 of Berry (upper Middle Ordovician) were obtained from the sulfidic black slate unit (Harwood and Berry, 1967). Rocks of similar character and age are widespread in the northern Appalachians (Berry, 1968).

Rocks of the upper part of division II are in conformable contact with rocks of division III. The transition is gradational and coincides with the first appearance of felsic metavolcanic rocks and thickly bedded metagraywacke of the Quimby Formation (Moench, 1969).

Upper Ordovician(?) rocks (division III). Division III includes the Quimby and Greenvale Cove Formations which are exposed at the south end of Kennebago Lake, at Rangeley Lake, and in the core of the Brimstone Mountain anticline to the south. Rocks of this division have not been recognized elsewhere in the region. The Quimby is composed of about 1000 feet of thickly bedded metagraywacke, conglomeratic metagraywacke, felsic metavolcanic rocks, and sulfidic black metashale overlain by about 2000 feet of thinly interbedded, sulfidic metashale and metagraywacke, sparse felsic metavolcanic rocks, and local conglomeratic metagraywacke. The upper few hundred feet of the formation are particularly sulfidic and carbonaceous and contain thin beds of dense black calcareous cherty rock. Clasts in the conglomeratic rocks are chert, felsic and mafic metavolcanic rocks, quartzite, and vein quartz. No clasts that were metamorphosed before erosion and deposition have been identified.

The Greenvale Cove Formation is a thin but distinctive and extensive unit that conformably overlies the Quimby Formation and underlies the Lower Silurian(?) Rangeley Formation (IV-1). The Greenvale Cove is composed of 300 to 600 feet of light-hued, noncarbonaceous, thinly interlaminated, fine-grained clastic rocks. The silty and sandy rocks are commonly calcareous, and in the southern outcrops they form a conspicuous unit of biotite granofels and calc-silicate rock that is similar to parts of the much younger Madrid Formation (IV-5b).

The Quimby and Greenvale Cove Formations are considered Late Ordovician(?) in age, on the basis of their stratigraphic position between fossil-bearing rocks of late Middle Ordovician and Early Silurian(?) age.

<u>Silurian and Silurian(?) rocks (lower part, division IV)</u>. These rocks include the Rangeley (IV-1), Perry Mountain (IV-3), Smalls Falls (IV-4) and Madrid Formations (IV-5b) (Moench, 1970; Osberg and others, 1968), and unnamed fossiliferous rocks in the Kennebago Lake quadrangle. As described later, probable correlatives of at least the Smalls Falls and Madrid Formations are recognized east of the Lexington batholith (IV-4, IV-5a, and IV-5b). In addition, Osberg and others (1968) have correlated the Rangeley, Perry Mountain, and Madrid Formations respectively with the Mayflower Hill, Waterville, and Vassalboro Formations exposed on the southeast limb of the Merrimack synclinorium. The Rangeley Formation is considered to be Early Silurian(?) in age, but fossiliferous rocks in the Kennebago Lake quadrangle, with which the Rangeley is correlated, are Early Silurian in age (see Trip A-1). Moench (1970) and Osberg and others (1968) have assigned a Silurian(?) age to the Perry Mountain, Smalls Falls, and Madrid Formations on the basis of correlations with fossiliferous rocks to the southeast, but recognize that the Madrid Formation may be Early Devonian in age. Silurian and Silurian(?) rocks are nearly 15,000 feet thick in the Phillips and Rangeley quadrangles. They thin drastically, however, across strike to the northwest. Facies relationships indicate a northerly or northwesterly source for at least the coarse clastic rocks of the Rangeley, Perry Mountain, and Smalls Falls Formations.

Near the east end of Rangeley Lake, the Rangeley Formation (IV-1) is nearly 10,000 feet thick and is characterized near its base by polymictic metaconglomerate with clasts of plutonic and volcanic rocks near the base, interbedded metashale and quartz-rich polymictic metaconglomerate in the middle, and interbedded metashale and quartz conglomerate near the top. To the north, near the south end of Kennebago Lake, the Rangeley thins greatly, but individual units of the formation are recognizable. The northern, coarse clastic facies of the Rangeley passes by intertonguing into a thick mass of finer-grained, gray metashale southward. The Taconic unconformity is not recognized in this area because the contact between the Greenvale Cove and Rangeley Formations is conformable and gradational. The contact between the Rangeley and Perry Mountain Formations is gradational over several tens of feet.

The Perry Mountain Formation (IV-3) is composed of nearly 2000 feet of mature, quartz-rich metasandstone cyclically interbedded with roughly equal amounts of light-hued muscovite-rich metashale. The formation is typically thin bedded, but beds of metasandstone and quartz-granule conglomerate as much as 10 feet thick are common in the middle and upper parts of the formation in the northern part of the main strike belt. Crossbedding, convolute bedding, and graded bedding are characteristic. Petrographic and chemical analyses indicate that the metasandstones are generally richer in silica than those of older and younger rocks and that the metashales are somewhat more potassic and aluminous than other metashales of the entire stratigraphic sequence. These relationships suggest that the inferred source area to the north was undergoing saprolitic weathering and shed chemically differentiated clastic rocks. The contact between the Perry Mountain and Smalls Falls Formations is sharp, conformable, and well

exposed in the Rangeley quadrangle and to the southwest. Depositional structures at the contact in both formations indicate that the Smalls Falls is younger.

The Smalls Falls Formation (IV-4) is a rusty-weathering unit composed of black sulfidic metashale cyclically interbedded with sulfidic quartz-rich metasandstone. The upper few hundred feet is calcareous. Thick graded beds of commonly calcareous quartz granule metaconglomerate are abundant in northern outcrops, particularly near the faults northwest of Madrid (Moench, 1970). The Smalls Falls Formation is about 2500 feet thick west and south of Madrid, but thins and wedges out northward. The contact between the Smalls Falls and Madrid Formations is sharp and conformable.

The Madrid Formation (IV-5b) is composed of approximately 1000 feet of calcareous metasiltstone and metasandstone with subordinate amounts of metashale. The formation is divisible into a lower part, composed of thin-bedded, calcareous clastic rocks, and an upper part composed of thick-bedded, fine-grained metasandstone.

Lower Devonian(?) rocks (upper part, division IV). Dominantly pelitic rocks of the Merrimack synclinorium of Early Devonian(?) age are widely exposed along the southeast margin of the area shown on Figure 1. In synclines east of Madrid and north of Kingfield, the Madrid Formation (IV-5b) is conformably overlain by an unknown thickness of medium-gray metashale, with minor amounts of metasandstone (IV-7h and IV-7s). The metashale, here in the staurolite grade of metamorphism, is variably massive, faintly cyclically bedded, and distinctively cyclically interbedded with light-gray or white beds of metasandstone. Although grouped with similar units here that are correlated with the Seboomook Formation (IV-7s), this massive metashale unit is also recognized in the Carrabassett valley as the lowermost member of a formational unit distinguished from the cyclically bedded, graded-bedded Seboomook, but correlated with the Lower Devonian(?) dominantly pelitic rocks of the region shown on Figure 1.

Southeast of the Blueberry Mountain fault is a group of rocks (IV-7h, IV-7s) called the southern sequence of Devonian(?) rocks by Moench (1970). This sequence is considered to be partly equivalent to rocks correlated with the Devonian(?) (IV-7s) northwest of the fault, but it contains a more varied lithic assemblage. In addition, the southern sequence has at least two prominent units of rusty-weathering, black sulfidic metashale; cyclically thin-bedded metashale and metasandstone similar to the Perry Mountain Formation (IV-3); and the Hildreths Formation (IV-7h, which is a thin but extensive unit of plagioclase- and biotite-rich metasandstone, calc-silicate rock, and local ribbon marble. The Hildreths and associated rocks are probably equivalent to similar rocks mapped in the upper part of the Lower Devonian sequence of the Carrabassett River valley north of Kingfield.

Silurian-Devonian correlation, Northwest flank of the Merrimack synclinorium to the southeast flank of the Moose River synclinorium

The stratigraphy and structural features of the Merrimack and Moose River synclinoria have been described (Osberg and others, 1968; Moench, 1970; Boucot, 1969), and therefore only the salient features of each that relate to fundamental regional correlation will be reviewed here. The regional map (fig. 1) extends northeastward primarily for this purpose, even though trip itineraries do not extend into the northeastern part of the map area.

Because of the scarcity of fossils in the Silurian and Devonian metamorphic rocks of the Rangeley, Phillips, northwest Kingfield, and Little Bigelow Mountain quadrangles, correlation along and across the regional strike has depended greatly on the calc-silicate-bearing rock units as markers in an otherwise uniform pelitic sequence. The lithologic sequence first established in the Phillips quadrangle, involving the Perry Mountain (IV-3), Smalls Falls (IV-4), and Madrid Formations (IV-5b), is a critical link to the stratigraphic sequence across the Sugarloaf gabbro pluton and Lexington batholith. Lack of outcrop prevents walking out the Madrid Formation from Salem (Phillips quadrangle) to Kingfield. There is little question, however, that remarkably similar lithic sequences at Kingfield are directly on strike with those near Salem and further northeastward east of the Lexington batholith. The Madrid has been mapped in detail to Wyman Lake and has been traced in reconnaissance farther northeastward.

The Smalls Falls Formation (IV-4), which directly underlies the Madrid, is not exposed in eastern Phillips quadrangle, nor in the Carrabassett valley area of the Little Bigelow Mountain quadrangle and northwestern part of the Kingfield quadrangle. What is believed to be Smalls Falls reappears southeast of the Lexington batholith. Brian D. Keith (unpublished manuscript) found top-facing sedimentary structures in the Smalls Falls and overlying Madrid equivalents in a well-exposed contact at the dam at the southeast end of Wyman Lake.

Northwest of the Salem anticline is a synclinal belt, complexly modified by intrusive-tectonic crowding, which extends to the Bigelow Mountain Range and northeastward beyond the Lexington batholith. In the northeastern area, massive metapelite grades into medium- to thinly bedded, graded-bedded metagraywacke and metapelite (all belonging to IV-7s) toward the axial zone of the synclinal belt. Where the axial zone is truncated by the Lexington batholith, infolded lenses of a younger, much more aluminous, sulfide-rich thinly layered calc-silicate and metapelite unit (IV-7h) occur.

Southeast of Pierce Pond and northeastward, G.M. Boone, and E.V. Post (unpublished data) have found southeast-facing sedimentary structures in gray massive Lower Devonian(?) slate (IV-7s) near the contact with underlying calcic metapelite (IV-5a). The calcic metapelite, forming the

northwest margin of the synclinal belt noted in the paragraph above, is much thinner, more aluminous, and much less quartzose than the Madrid Formation (IV-5b) but occupies a similar stratigraphic position with respect to the overlying Lower Devonian(?) metapelite (IV-7s). Similarly, along the southeast margin of the Moose River synclinorium, thin units of predominantly calcareous slate and limestone (IV-6) are found at the base of the Silurian-Devonian stratigraphic sequence (Boucot, 1961; 1969). These basal units are in fault contact (southeast boundary fault) with underlying rocks of division II exposed in the Dead River syncline and Hurricane Mountain anticline.

The position of the basal calcareous units (IV-5a and IV-6) with respect to the immediately underlying pre-Silurian section that they flank, contrasts sharply with the thick Silurian section underlying the Madrid Formation farther southwest. Although the two calcic horizons, on the northwest and southeast sides of the synclinal belt respectively, cannot be directly correlated with each other, they are probably of similar age, later Silurian to Early Devonian. Osberg and others (1968) favor a Late Silurian(?) for the Madrid Formation (IV-5b); E.V. Post (unpublished data) reports a Wenlock to Ludlow age for fossils found in a carbonate-rich part of IV-5a. Shelly fauna from the IV-5a in the Little Bigelow Mountain quadrangle is less definitive but typical of the Silurian (A.J. Boucot, written commun., 1967, 1969). Fauna in calc-silicate rock of IV-1 in the Stratton quadrangle is dated as upper Llandoverian (Boucot, 1969).

The absence of a thick Silurian section below the thin carbonate-bearing and calc-silicate units in the central and northeastern part of the synclinal belt east of the Lexington batholith implies that the tectonic hinge formed a major control on the pattern of sediment deposition and affected lithic distribution and thickness throughout the Silurian. Varied calcic horizons in the Devonian rocks in the Merrimack synclinorium (not differentiated in Figure 1) may, therefore, project to the same apparent stratigraphic position below Lower Devonian metapelites (IV-7s) flanking the older depositional prism and the Hurricane Mountain anticline containing rocks of division II.

Correlations in the southeast margin of the Connecticut Valley-Gaspé synclinorium and the Moose River synclinorium

Lower Paleozoic eugeosynclinal rocks, undivided. Undifferentiated lower Paleozoic eugeosynclinal rocks (Pz), which include gray to interbedded gray and silver-green slate and phyllite, dark metasandstone and sparse metagraywacke, felsic metavolcanic rocks, and mafic metavolcanic rocks with pillow structure, compose the sequence here assigned to the Connecticut Valley-Gaspé synclinorium. Great variation in thickness of bedding is characteristic of these rocks, and grading and crosslamination are rarely observed. The sequence is probably more than 10,000 feet thick in western Maine. This lithologic unit was correlated with the Frontenac Formation (McGerrigle, 1935) by Marleau (1968), but Marleau thought the unit was younger than rocks mapped as the Seboomook Formation (IV-7s) of Early Devonian age or those correlated with it (Compton Formation in Québec). Boucot (1961) extended the name Frontenac Formation into Maine for rocks

on strike (Pz near Jackman), and arbitrarily correlated the unit across regional strike with the Tarratine Formation (IV-8) of Early Devonian age. The Pz unit bears a close similarity to rocks mapped beneath the Seboomook Formation in the St. John-Allagash River basins (Boudette and others, 1967), and boundary relationships between Pz and the Seboomook Formation in the northern part of the region suggest that a fault of significant magnitude separates them and transports the unit up and over the Seboomook (E.L. Boudette unpublished data). Thus its age remains uncertain (see Boucot, 1969, p. 43-46). We have provisionally shown Pz as Silurian(?) or Ordovician(?). No obvious lithologic correlatives exist southeast of the Boundary Mountains anticlinorium.

Lower Devonian rocks

The youngest metasedimentary units in the northwest and northeast part of the region are, in ascending order, the Seboomook (IV-7s), Tarratine (IV-8), and Tomhegan (IV-9) Formations (Boucot, 1961, 1969). The Tarratine and Tomhegan underlie the axial area and northwestern flank of the Moose River synclinorium in the northeast.

The Seboomook, described above, is as much as 15,000 feet thick and is dated by fossils as of Early Devonian age (Boucot, 1969, p. 35).

The Tarratine Formation is composed of interbedded dark metasandstone and metapelite with subordinate quartzite near the top and metalimestone near the bottom of the unit; its total thickness is about 10,000 feet. Bedding thickness of great variability from fractions of an inch to 50 feet characterize the unit which is dated by fossils as of Early Devonian age (Boucot, 1969, p. 27-28).

The Tomhegan Formation is composed mainly of thickbedded, dark metasandstone interbedded with thin and thick-bedded metapelite in the upper part of the unit, and felsic to intermediate metavolcanic rocks in the lower part. The unit totals about 4,000 feet in thickness and is ascribed a paleontologic age of Schoharie and possible Oriskany (Boucot, 1969, p. 20). The contact between Tarratine and Tomhegan is structurally conformable, but a disconformity is indicated by a faunal break (Boucot, 1969, p. 19-20). The Tarratine is in gradational contact with the Seboomook Formation (IV-7s) or lies unconformably upon the Attean Quartz Monzonite (unit 2).

Intrusive rocks.

Upper Cambrian(?) or Lower Ordovician(?) ultramafic complex. The complex, now metamorphosed, consists of medium- to coarse-grained epidiorite, epidiorite autobreccia, serpentinite, and altered quartz diorite with minor amounts of altered pyroxenite, gabbro, and alaskite. The complex is a differentiated sheet intruded essentially along the pre-Lower Ordovician unconformity; distinctive igneous layering is produced locally in the epidiorite facies. Serpentinite can be divided into contiguous (antigorite-bearing) and diapiric (chrysotile-bearing) types. Rocks of the complex have not been dated by radiometric methods, but the Attean Quartz Monzonite, which may be a part of the complex, is provisionally dated by the Pb-U and

other methods (see below). Because the Attean cuts the ultramafic complex, the complex is older but is believed to be almost time equivalent. Both the rocks of the ultramafic complex and the Attean intrude the Chain Lakes massif and, hence, the complex is younger than the Chain Lakes and is provisionally considered to be of Late Cambrian to Early Ordovician age by correlation with similar rocks in Quebec (St. Julien, 1967).

Lower Ordovician(?) quartz monzonite and granodiorite. These units are largely leucocratic, coarse-grained, porphyritic, characterized by alteration, and locally deformed. Plutons assigned to this group include the Attean Quartz Monzonite (Albee and Boudette, in press) and the Parmachenee Granodiorite, Adamstown Granite, and Umbagog Granodiorite of Green and Guidotti (1968). The Attean has been dated by the Pb-U method; giving a radiometric age of ca. 470 m.y. (Leon T. Silver, personal commun., 1970). Because the Attean quartz monzonite has not been found intruding Middle Ordovician rocks, we consider the radiometric age indicates a probable Lower Ordovician age according to the Holmes Scale.

Quartz porphyry and related felsic rocks. Dikes and irregular bodies of quartz porphyry, granophyre, and fine- to medium-grained quartz monzonite intrude the Attean Quartz Monzonite and ultramafic complex in numerous places. The most distinctive of these localities is at Catheart Mountain in the Long Pond quadrangle where a Mo-Cu sulfide deposit is related to the porphyry. Radiometric dating of white mica in greisen associated with the porphyry at Catheart Mountain indicates that the age of the rock is between 433 m.y. and 457 m.y. (F.C. Canney, unpublished data). Textures within the quartz porphyry are suggestive of hypabyssal emplacement in an event that would require the Attean Quartz Monzonite and other host rocks to have been uplifted and exposed or nearly exposed after crystallization. Regional paleogeographic considerations suggest that such an environment was most likely in Late Ordovician or Early Silurian time. We provisionally assign the hypabyssal series to the Late Ordovician.

Lower Devonian garnet rhyolite and related felsic rocks. Hypabyssal sills, dikes, and small stocks of light-gray, fine-grained massive garnet rhyolite and felsic hypabyssal rocks cut the Tarratine Formation (IV-8) in the Moose River synclinorium. These plutonic rocks are petrochemically and spatially correlated with volcanic rocks in the Tomhegan Formation (IV-9) (Boucot, 1969, p. 59-60).

Devonian (syn- and post-tectonic) mafic intrusive rocks. A segmented, but regionally conformable belt of mafic intrusive rocks extends from the Kennebago Lake and Rangeley quadrangles northeastward beyond the conference region to Moosehead Lake and eastward (see Hussey and others, 1967). Norite, troctolite, gabbro, diorite, quartz diorite, and related rocks occur within plutons of the belt which intrude rocks as young as the Lower Devonian Seboomook Formation (IV-7s). All these rocks are associated with well-developed, commonly spectacular contact aureoles impressed upon the Acadian or older regional metamorphic assemblages which seem to have slightly predated the development of the hornfels rocks of the aureoles.

Reaction and injection hornfelses of several varieties are common, and layering comparable to that in the older ultramafic complex is not present, although flow structure and igneous layering is observed locally (Espenshade and Boudette, 1967, p. F-34). Norite near Greenville, Maine, has been dated radiometrically (biotite, K-Ar method) at 393 m.y. by Faul and others (1963). Thus, the entire belt of mafic plutons is interpreted to be Early Devonian in age.

Devonian (posttectonic) quartz monzonite and granodiorite. Widespread granitic plutons of varied size discordantly intrude metasedimentary units as young as the Seboomook Formation (IV-7s) as well as igneous rocks as young as the Devonian mafic rocks. These plutons are largely leucocratic varieties of coarse- to fine-grained quartz monzonite and granodiorite which vary from porphyritic to equigranular. These rocks have also produced well-defined contact aureoles similar to those adjacent to the Devonian mafic plutons. Radiometric dating (see Faul and others, 1963; Lyons and Faul, 1968) of the Hog Island Granodiorite near Jackman Station (Albee and Boudette, in press), gives an age of 360 m.y., but similar rocks emplaced in belts on strike to the south give slightly greater ages, to about 380 m.y. (see Faul and others, 1963, p. 4). The intrusive relationships and the radiometric dating suggest that although the granitic rocks may be younger in age than the Devonian mafic intrusives they are post-Acadian, and may also be late Early Devonian or possibly early Middle Devonian in age. Comagmatic relationships, however, are not implied.

Triassic(?) lamprophyre dikes and sills. Dark lamprophyre dikes and sills are widespread throughout the region. These rocks are nowhere abundant but apparently cut all the older rocks. The lamprophyres occur in markedly tabular bodies, ranging from a few feet to a few tens of feet thick, which usually show distinct chilled margins. Textures and modes of the lamprophyre rocks are variable, but the rocks are commonly porphyritic varieties. None of these rocks has been dated by radiometric methods. Our assignment of a Triassic(?) age to them is arbitrary.

REFERENCES CITED

Albee, A.L., 1961, Boundary Mountain [sic] anticlinorium, west-central Maine and northern New Hampshire: U.S. Geol. Survey Prof. Paper 424-C, p. C51-C54.

Albee, A.L., and Boudette, E.L., 1971, Geology of the Attean quadrangle, Somerset County, Maine: U.S. Geol. Survey Bull. 1297, in press.

Berry, W.B.N., 1968, Ordovician paleogeography of New England and adjacent areas based on graptolites, in Zen, E-an White, W.S., Hadley, J.B., and Thompson, J.B., Jr., editors, Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 23-34.

Billings, M.P., 1956, The geology of New Hampshire; Part II, Bedrock geology: Concord, N.H., New Hampshire State Plan. and Devel. Comm., 203 p.

Boone, G.M., in prep., Stratigraphy and metamorphism in the Little Bigelow Mountain quadrangle, Maine: Maine Geol. Survey Bull.

Boucot, A.J., 1961, Stratigraphy of the Moose River synclinorium, Maine: U.S. Geol. Survey Bull. 1111-E., p. 153-188.

_______ 1969, Geology of the Moose River and Roach River synclinoria, northwestern Maine: Maine Geol. Survey Bull. 21, 117 p.

Boucot, A.J., and Thompson, J.B., Jr., 1963, Metamorphosed Silurian brachiopods from New Hampshire: Geol. Soc. America Bull., v. 74, no. 11, p. 1313-1333.

Boucot, A.J., Griscom, Andrew, and Allingham, J.W., 1964, Geologic aeromagnetic map of northern Maine: U.S. Geol. Survey Geophys. Inv. Map GP-312.

Boudette, E.L., Hatch, N.L., Jr., and Harwood, D.S., 1967, Geology of the upper St. John and Allagash River basins, Maine: U.S. Geol. Survey open-file rept., 77 p.

Boynton, G.R., and Gilbert, F.P., 1964a, Aeromagnetic map of the Cupsuptic quadrangle, Oxford and Franklin Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-477.

_______ 1964b, Aeromagnetic map of the Oquossoc quadrangle, Oxford and Franklin Counties, Maine: U.S. Geol. Survey Geophys. Map GP-478.

_______ 1964c, Aeromagnetic map of the Phillips quadrangle, Franklin County, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-479.

_______ 1964d, Aeromagnetic map of the Rangeley quadrangle and part of the Kennebago Lake quadrangle, Franklin and Oxford Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-480.

Bromery, R.W., Kirby, J.R., Vargo, J.L., and others, 1957a, Aeromagnetic map of Umbagog Lake and vicinity, New Hampshire: U.S. Geol. Survey Geophys. Inv. Map GP-138.

_______1957b, Aeromagnetic map of Berlin and vicinity, New Hampshire: U.S. Geol. Survey Geophys. Inv. Map GP-139.

Bromery, R.W., and Gilbert, F.P., 1962, Aeromagnetic map of the Skinner and parts of the Attean and Sandy Bay quadrangles, Somerset and Franklin Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-310.

✓ Bromery, R.W., Soday, Harry, and others, 1963, Aeromagnetic map of the Spencer Lake quadrangle, Franklin and Somerset Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-331.

Bromery, R.W., Tyson, N.S., and others, 1963, Aeromagnetic map of the Stratton quadrangle, Franklin and Somerset Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-333.

Cady, W.M., 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, no. 5, p. 531-576.

Childs, J.F., and Boone, G.M., 1968, Metamorphism of pre-Silurian chloritic phyllite to sillimanitic granofels, west-central Maine: [abs.] Geol. Soc. America Spec. Paper 101 p. 252.

Espenshade, G.H., and Boudette, E.L., 1967, Geology and petrology of the Greenville quadrangle, Piscataquis and Somerset Counties, Maine: U.S. Geol. Survey Bull. 1241-F, 60 p.

Evans, B.W., and Guidotti, C.V., 1966, The sillimanite-potash feldspar isograd in western Maine, U.S.A.: Contr. Mineralogy and Petrology, v. 12, no. 1, p. 25-62.

Faul, Henry, Stern, T.W., Thomas, H.H., and Elmore, P.L.D., 1963, Ages of intrusion and metamorphism in the northern Appalachians: Am. Jour. Sci., v. 261, no. 1, p. 1-19.

Green, J.C., 1963, High-level metamorphism of pelitic rocks in northern New Hampshire: Am. Mineralogist, v. 48, nos. 9-10, p. 991-1023.

✓ _______1964, Stratigraphy and structure of the Boundary Mountain anticlinorium in the Errol quadrangle, New Hampshire-Maine: Geol. Soc. America Spec. Paper 77, 78 p.

✓ _______1968, Geology of the Connecticut Lakes-Parmachenee area, New Hampshire and Maine: Geol. Soc. America Bull., v. 79, no. 11, p. 1601-1638.

Green, J.C., and Guidotti, C.V., 1968, The Boundary Mountains anticlinorium in northern New Hampshire and northwestern Maine, in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B., Jr., editors, Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 255-266.

Harwood, D.S., 1966, Geology of the Cupsuptic quadrangle, Maine: U.S. Geol. Survey open-file rept., 259 p.

_______1968, Geology of the Arnold Pond quadrangle, Oxford and Franklin Counties, Maine: U.S. Geol. Survey open-file rept., 13 p.

_______1969, The Second Lake anticline--A major structure on the northwest limb of the Boundary Mountain [sic] anticlinorium, northern New Hampshire, west-central Maine, and adjacent Quebec: U.S. Geol. Survey Prof. Paper 650-D, p. D106-D115.

Harwood, D.S., and Berry, W.B.N., 1967, Fossiliferous lower Paleozoic rocks in the Cupsuptic quadrangle, west-central Maine: U.S. Geol. Survey Prof. Paper 575-D, p. D16-D23.

Henderson, J.R., Gilbert, F.P., and others, 1963a, Aeromagnetic map of the Chain Lakes quadrangle, Franklin and Somerset Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-330.

_______1963b, Aeromagnetic map of the Kennebago Lake quadrangle, Franklin County, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-332.

Henderson, J.R., and Smith, C.W., 1964, Aeromagnetic map of the Old Speck Mountain quadrangle, Franklin and Oxford Counties, Maine: U.S. Geol. Survey Geophys. Inv. Map GP-486.

Hurley, P.M., and Thompson, J.B., Jr., 1950, Airborne magnetometer and and geological reconnaissance survey in northwestern Maine: Geol. Soc. America Bull., v. 61, no. 8, p. 835-841.

Hussey, A.M., Chapman, C.A., Doyle, R.G., Osberg, P.H., Pavlides, Louis, and Warner, Jeffrey, compilers, 1967, Preliminary geologic map of Maine: Augusta, Maine Geol. Survey, scale 1:500,000.

Lyons, J.B., and Faul, Henry, 1968, Isotope geochronology of the northern Appalachians, in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B., Jr., editors, Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 305-318.

Marleau, R.A., 1968, Woburn-East Mégantic-Armstrong area, Frontenac and Beauce Counties [Quebec]: Quebec Dept. Nat. Resources Geol. Rept. 131, 55 p.

McGerrigle, H.W., 1935, Mount Mégantic area, southeastern Québec, and its placer-gold deposits: Québec Bur. Mines Ann. Rept. 1934, pt. D, p. 63-104.

Moench, R.H., 1963, Prelimary geologic map of the Phillips quadrangle, Maine: U.S. Geol. Survey Mineral Inv. Field Studies Map, MF-259.

_______1966, Relation of S_2 schistosity to metamorphosed clastic dikes, Rangeley-Phillips area, Maine: Geol. Soc. America Bull., v. 77, no. 12, p. 1449-1461.

_______1969, The Quimby and Greenvale Cove Formations in western Maine: U.S. Geol. Survey Bull. 1274-L, 17 p.

_______1970, Prematamorphic down-to-basin faulting, folding, and tectonic dewatering, Rangeley area, western Maine: Geol. Soc. America Bull. v. 81, no. 5, p. 1463-1496.

Osberg, P.H., Moench, R.H., and Warner, Jeffrey, 1968, Stratigraphy of the Merrimack synclinorium in west-central Maine, in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B., Jr., editors, Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 241-253.

Pavlides, Louis, Boucot, A.J., and Skidmore, W.B., 1968, Stratigraphic evidence for the Taconic orogeny in the northern Appalachians, in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B., Jr., editors, Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 61-82.

Poole, W.H., Béland, J.R., and Wanless, R.K., 1963, Minimum age of Middle Ordovician rocks in southern Québec: Geol. Soc. America Bull., v. 74, no. 8, p. 1063-1065.

St. Julien, Pierre, 1967, Tectonics of part of the Appalachian region of southeastern Québec (southwest of the Chaudière River): Royal Soc. Canada Spec. Pub. 10, p. 41-47.

U.S. Geological Survey, 1969, Aeromagnetic map of the First Connecticut Lake and vicinity, northern New Hampshire: U.S. Geol. Survey Geophys. Inv. Map, GP-654.

Winkler, H.G.F., 1967, Petrogenesis of metamorphic rocks (2d ed.): New York, Springer-Verlag, Inc., 237 p.

Introduction to the Quaternary history in the highlands region of western Maine, southeastern Québec, and northern New Hampshire

Harold W. Borns, Jr., Parker E. Calkin, Carl Koteff[2], Fred Pessl[2], and W.W. Shilts[3].

(The authorship of each section is indicated in the text of the introduction.)

The highlands of northwestern Maine, including the Longfellow and Boundary Mountains, were overridden at least twice and perhaps several more times, by continental ice sheets during the Quaternary Period. These episodes are indicated by five, widely separated exposures displaying two-drifts sequences composed of lodgment tills separated by lacustrine and fluvial sediments. The freshness of these drifts suggest that they are probably of Wisconsin age, however this is equivocal as no way has been found to assign absolute ages to them.

Caldwell (1959) reports a two-till sequence at New Sharon, in central Maine, separated by organic materials dated at more than 38,000 years old. A recent C^{14} age determination (Stuiver, personal communication), and an analysis of the wood fragment which shows that the tree was crushed while still green, indicate that the organic material at New Sharon was overridden by ice more than 52,000 years ago. Presently there is no way of determining the relationship of the New Sharon sequence and the undated sequences to the northwest.

The last ice sheet, whose retreating margin stood along the present Maine coast approximately 13,500 years ago (Borns, unpublished), thinned, separated and stagnated over the Longfellow and Boundary Mountains of northwestern Maine, a belt 60 miles wide and rising over 3000 ft. above bordering lowlands. Nearly contemporaneous stagnation, throughout and perhaps to the southeast of the mountains, is evidenced by the distribution and volume of ice-contact stratified drift. Coupled with this is the lack of evidence of a receding active ice margin.

The separation of this ice in Maine from the still-active receding ice sheet immediately to the northwest in Québec, occurred approximately 12,800 years ago and the subsequent dissipation of stagnant ice in the mountains was complete by approximately 12,000 years ago.

---

[2]Publication authorized by the Director, U.S. Geological Survey.

[3]Publication authorized by the Director, Geological Survey of Canada.

The highest glacial cirques in northwestern Maine on Crocker Mtn., with floors at an altitude of approximately 2700 ft., reveal no evidence of glacial reactivation during and subsequent to the dissipation of the last ice sheet.

Harold W. Borns, Jr.
University of Maine

Parker E. Calkin
S.U.N.Y. at Buffalo

Tills of three glaciations are exposed in the Lac-Mégantic area; non-glacial fluvial sediments bearing reworked erratics from the Canadian Shield underlie the oldest till at the Grande Coulée River and indicate erosion of a terrane covered by a fourth, earlier glacial deposit. The three tills, from oldest to youngest, are named Johnville, Chaudière, and Lennoxville Tills.

Johnville Till was deposited by a glacier flowing southeast; it is overlain by the Massawippi Formation, composed of non-glacial sediments dated at >40,000 BP (GSC-1084). The Massawippi Formation contains pollen indicating climate colder than present and is tentatively correlated with the St. Pierre sediments of the St. Lawrence Lowlands.

Chaudière Till overlies Massawippi sediments. It was deposited by a glacier which first flowed southwest into and over the Lac-Mégantic region. At some time during the Chaudière glacial phase ice-flow shifted to southeast. Ice-flow directions are confirmed by fabric and petrographic data.

The Chaudière glacier is inferred to have retreated only to the St. Lawrence Lowlands. As it rested against the west- and north-facing slopes of the Appalachians, it blocked northward drainage and ponded a large lake in the Chaudière and St. Francis river basins. The lake is named Glacial Lake Gayhurst and sediments deposited in it comprise the Gayhurst Formation.

The Gayhurst Formation, consisting of a lower member of about 3400 graded silt-clay laminae, a middle member of shallow water sand and gravel, and an upper member of about 600 graded silt-clay laminae was deposited in a 380 m outlet phase (lower and middle members) and a 430 m outlet phase (upper member) of Glacial Lake Gayhurst. The 380 m outlet carried overflow east into the St. John River; the 430 m outlet carried overflow southeast through Coburn Gore, Maine and operated whenever ice stood far enough south of the Boundary Mountains to block the 380 m outlet. It is suggested that the Gayhurst Formation was deposited during the entire time interval separating deposition of Chaudière Till and overlying Lennoxville Till; 4000 couplets of laminated silt-clay may represent at least 4000 years of deposition. Presumed Gayhurst Formation sediments have been dated at >20,000 B.P. (GSC-1137).

Lennoxville Till was deposited during a major readvance over Gayhurst Formation sediments. The Lennoxville glacier is inferred to have covered all of southeastern Québec and New England. Striae, fabric, and indicator dispersal patterns indicate east-southeast movement of the Lennoxville glacier at its maximum. The Drolet Lentil of Lennoxville Till, partially derived from the clayey sediments of the Gayhurst Formation, was deposited during early south-southwest Lennoxville glacier advance up the Chaudiere valley.

The Lennoxville glacier retreated from the Mégantic area by back-wasting of actively-flowing ice. During halts or readvances the glacier built till and gravel moraines and impounded proglacial lakes in the Chaudière valley and its tributaries. As the Lennoxville glacier melted, large boulders carried in and on the ice were let down onto the surface of Lennoxville lodgment till as a one-boulder-thick ablation deposit.

No evidence was found to support the concept of late-glacial ice flow into Québec from highland centers in Maine or New Hampshire.

William W. Shilts
Geological Survey of Canada
Ottawa, Canada

Recently available exposures on Nash Stream in northern New Hampshire are the most informative to date that bear on the major controversy over the relative ages of two widespread tills in New England. This controversy considers whether the two tills are contemporaneous deposits of a single ice sheet, or are deposits of two separate glaciations.

The exposures at Nash Stream show a composite section of a lower till as much as 70 feet thick, overlain by lacustrine sediments as much as 75 feet thick. These sediments are unconformably overlain by more than 100 feet of an upper till which in turn is overlain by outwash as much as 25 feet thick. The lower till is dark olive gray, compact, silty, and has subhorizontal and subvertical joints. It is oxidized locally to a depth of about 20 feet. The depth and nature of the oxidation of the lower till is nearly identical to that in similar tills throughout southern New England, and the oxidation is believed to represent a significant weathering interval between two glaciations. The lower till is not oxidized where overlain by the lacustrine sediments, strongly suggesting that the lacustrine beds are melt-water deposits of the retreatal phase of the lower-till ice sheet.

The upper till is light olive gray to olive, compact to friable, and shows no appreciable oxidation by weathering. This till is similar

to the upper till identified elsewhere in New England. At Nash Stream, the upper till truncates the lower lacustrine sediments and locally has well developed deformation (shove?) structures oriented southward. The upper outwash locally intertongues with the upper till and is therefore related in age to the younger ice sheet. At one locality, noncollapsed upper outwash truncates collapsed beds of lower lacustrine sand and gravel.

This sequence of deeply oxidized lower till associated with collapsed water-laid sediments, overlain by very thick nonoxidized till and associated outwash represents an outstanding set of exposures in New England that support the hypothesis of two major ice advances separated by a significant weathering interval.

Exposures of two superposed tills, similar to the tills along Nash Stream, are known in Connecticut, eastern Massachusetts, and southern New Hampshire. In the absence of dateable materials from these tills, regional correlation, even when restricted to southern New England, remains tentative. Extending the regional interpretation to include Maine and southern Québec is even more tenuous. However, the increasing body of field data seems to indicate that a widespread two-till stratigraphy exists throughout much of the northeast. Furthermore, the interpretation that the lower till may be pre-Sangamon in age, rather than early Wisconsinan (pre-classical Wisconsin), seems to be gaining favor. Whether more than two stratigraphically significant tills, as in southeastern Québec, are represented in interior New England remains unclear.

Several miles north of the well established stratigraphy, logs collected from till-like material are dated about 8500 years B.P. Dates from peat in till-like material stratigraphically higher than the logs are about 8300 years B.P. However, it is suggested that the logs are in landslide material composed primarily of reworked till and do not represent an ice advance at such a late date in New England.

Fred Pessl and Carl Koteff
U.S. Geological Survey
Boston, Massachusetts

TRIP A-1

Stratigraphy of the northwest limb of the Merrimack synclinorium in the Kennebago Lake, Rangeley, and Phillips quadrangles, western Maine[1]

by

R. H. Moench and E. L. Boudette
U.S. Geological Survey
Denver, Colorado and Boston, Massachusetts

Trip A-1 spans a major part of the northwest limb of the Merrimack synclinorium in western Maine. (See fig. 1 of Boone and others, this volume.) Except where faulted, the exposed sequence of Ordovician, Silurian, and Devonian rocks is continuous in most of the trip area, and is far thicker and more varied than any other recognized sequence along the same limb of the synclinorium to the southwest in New England. Despite tight folding and metamorphism to greenschist to amphibolite facies, sedimentary features are well preserved and have permitted detailed stratigraphic subdivisions and geologic mapping (fig. 1). In addition, the discoveries of late Middle Ordovician and Early Silurian fossils at the localities shown in the Cupsuptic and Kennebago Lake quadrangles (fig. 1; Harwood and Berry, 1967; U.S. Geological Survey, 1965, p. A-74) have permitted assignment of tentative ages to most of the exposed sequence in the Rangeley and Phillips quadrangles. The purpose of this trip is to examine the stratigraphic sequence from the Quimby Formation, of Late Ordovician(?) age, to the lower part of the Seboomook(?) Formation, of Devonian age, and to study the basis for correlating fossiliferous Lower Silurian quartz conglomerates in the Kennebago Lake quadrangle with quartz metaconglomerates of the part C of the Rangeley Formation of the Rangeley quadrangle. This correlation, if valid, has an important bearing on the stratigraphy of other parts of the Merrimack synclinorium, particularly in central and southern New England.

Because the purpose of this trip is stratigraphic, the terms metashale, metagraywacke, metasandstone, and so on, for convenience are used instead of their metamorphic equivalents. Metamorphic recrystallization alters primary textures, but primary structures are well preserved, except in calc-silicate rocks. The term "metagraywacke" is used here for unsorted sandstone composed of angular fragments of feldspar, rock fragments, and quartz set in an argillaceous matrix. Metagraywacke within the lowest metamorphic grades displays primary textural features. All the virtually noncalcareous metagraywackes and the more quartzose metasandstones have simple mineral

---

[1]Publication authorized by the Director, U.S. Geological Survey.

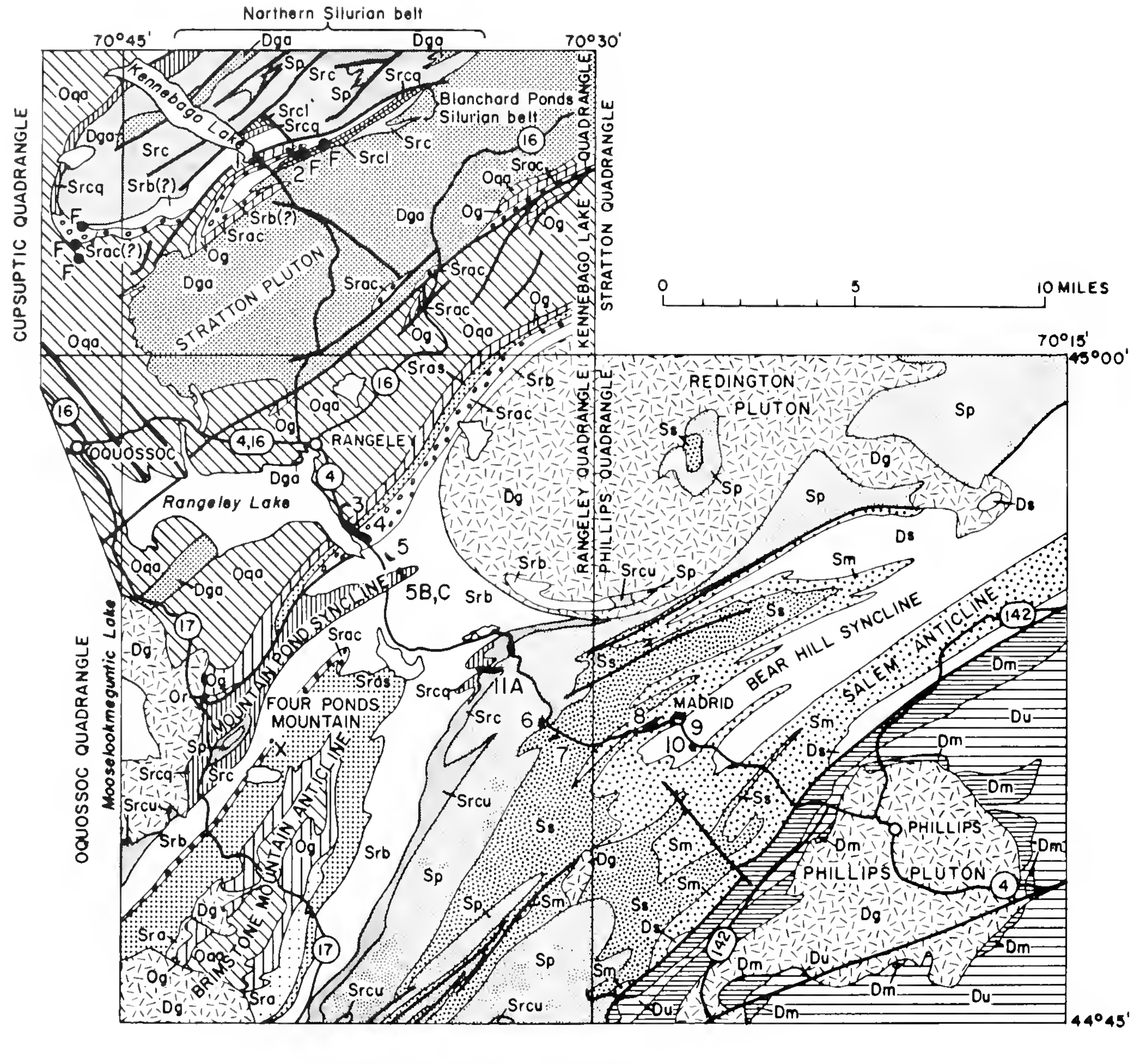


Figure 1.--Geologic map showing stops of trip A-1. Compiled by R.H. Moench and E.L. Boudette from Moench (1970); E.L. Boudette, unpublished mapping; and from modifications of Harwood and Berry (1967, fig. 2) and Harwood (1966).

EXPLANATION

DEVONIAN

Dg — Granitic rocks

Dgo — Gabbro, diorite, and associated reaction hornfels

Southern sequence

DEVONIAN(?)

Du — Upper unit
Nonsulfidic and sulfidic metashale

Dm — Middle unit
Metasandstone, calc-silicate rock, and marble of Hildreths Formation at top; nonsulfidic and sulfidic metashale

Lower Devonian — **DEVONIAN**

Ds — Seboomook(?) Formation
Dominantly gray metashale

Lower Silurian?

Sm — Madrid Formation
Calcareous metasandstone, metasiltstone, and calc-silicate rock

Ss — Smalls Falls Formation
Sulfide-rich black metashale and metasandstone; calcareous in upper part; grit abundant in north

Sp — Perry Mountain Formation
Cyclically interbedded commonly cross-laminated quartz-rich metasandstone and muscovite-rich metashale

Northern sequence

Rangeley Formation — **SILURIAN(?)**

Src, Srcl, Srcq, Srcu — Part C

Src, *gray metashale and subordinate feldspathic metasandstone*

Srcl, *laminated calcareous metasiltstone or silty marble. Absent in Rangeley quadrangle*

Srcq, *quartzite, quartz conglomerate, and fossiliferous calcareous quartz conglomerate and metasandstone in Kennebago Lake quadrangle; interbedded calcareous quartz conglomerate, feldspathic metasandstone, and gray metashale in Rangeley quadrangle*

F, *approximate location of fossil localities in Kennebago Lake quadrangle*

Srb — Part B
Gray metashale interbedded with subordinate feldspathic metasandstone; local quartz-rich polymictic metaconglomerate and conglomeratic mudflow deposits. Thin or absent in Kennebago Lake quadrangle

Srac, Sras, Sra — Part A

Srac, *polymictic metaconglomerate*

Sras, *feldspathic metasandstone*

Sra, *gray metashale and subordinate feldspathic metasandstone*

Upper Ordovician(?) — **ORDOVICIAN(?)**

Og — Greenvale Cove Formation
Interlaminated feldspathic metasandstone, metasiltstone, and metashale; commonly calcareous

Middle Ordovician and Upper Ordovician(?) — **ORDOVICIAN AND ORDOVICIAN(?)**

Oqa — Quimby, Dixville(?), and Albee Formations

1 5B,C 11A

Field-trip stops
B, C, *bus sections* B *and* C
A, *bus section* A

OF
Fossil locality

——?——
Contact
Queried where conjectural

Fault

assemblages, such as quartz, plagioclase (typically untwinned oligoclase), biotite (or chlorite and white mica at lowest grades), and locally sparse amounts of garnet. The mineralogy of the metashales in the southwestern part of the area is discussed in more detail by Guidotti (trip B-2, this volume). For the sake of simplicity, metamorphic zones are not shown on Figure 1. Stops 1 and 2 are in the contact metamorphic halo of the Stratton pluton; pelitic rocks here contain andalusite and cordierite. Stops 3 and 4 are in the biotite and garnet zones, respectively, and stops 5 through 11 are within a broad zone of staurolite-grade metamorphism. Most of the staurolite, however, has been retrograded to chlorite and sericite.

Because of the importance of the Rangeley Formation, our interpretations of its stratigraphic and facies relationships are briefly reviewed here and are schematically illustrated in Figure 2. Column 1 was constructed from sparse outcrops in the broad, grossly synclinal northern belt of Silurian(?) rocks that is crossed by Kennebago Lake (fig. 1). Silurian rocks of this belt typically have rather gentle dips. At the southwest end of the belt, steeply dipping upper Middle Ordovician graptolite-bearing slate is apparently unconformably overlain by less deformed clastic rocks of the Rangeley Formation (Harwood and Berry, 1967; this report, fig. 1). Column 2, constructed from rocks exposed at stop 2, represents a narrow belt of rocks between the southeast end of Kennebago Lake and the Stratton pluton (fig. 1). For convenience, this belt is called the Blanchard Ponds belt of Silurian(?) rocks. It is separated from the broad belt to the north by a fault of unknown displacement. Rocks of this belt strike northeast and dip nearly vertically. Tops of beds at stop 2 are interpreted to face southeast from sedimentary features. Columns 3, 4, and 5 (fig. 2) represent the main mass of the Rangeley Formation in the Rangeley quadrangle. Column 3 was constructed from data obtained from outcrops at and near stops 4, 5, and 5B, C (for trip sections B and C) southeast of Rangeley Lake, and outcrops north of the small canoe-shaped body of Perry Mountain Formation to the southwest (fig. 1). Column 4 represents a line of outcrops that entends west along a ridge from Four Ponds Mountain to the same canoe-shaped body of Perry Mountain. Column 5 represents several lines of outcrops near the southern border of the Rangeley quadrangle.

Rocks of the main mass of the Rangeley Formation in the Rangeley quadrangle are tightly folded, but are predominantly unfaulted. Correlation among columns 3, 4, and 5 is not a problem because the Rangeley Formation, as a whole, is sandwiched between the distinctive underlying Greenvale Cove and overlying Perry Mountain Formations, and because facies changes were mapped within this coherent mass of the Rangeley Formation. Correlations among columns 1, 2, and the main mass are inferred from similarities of rock types and sequence, which are consistent throughout, despite the obvious thickness changes (Fig. 2). The change in apparent thickness of parts A and B of the Rangeley between the areas of columns 2 and 3 amount to about 1-1/2 miles in a horizontal distance of 10 miles. This abrupt thickening suggest that the limb of the sedimentary basin has been shortened.

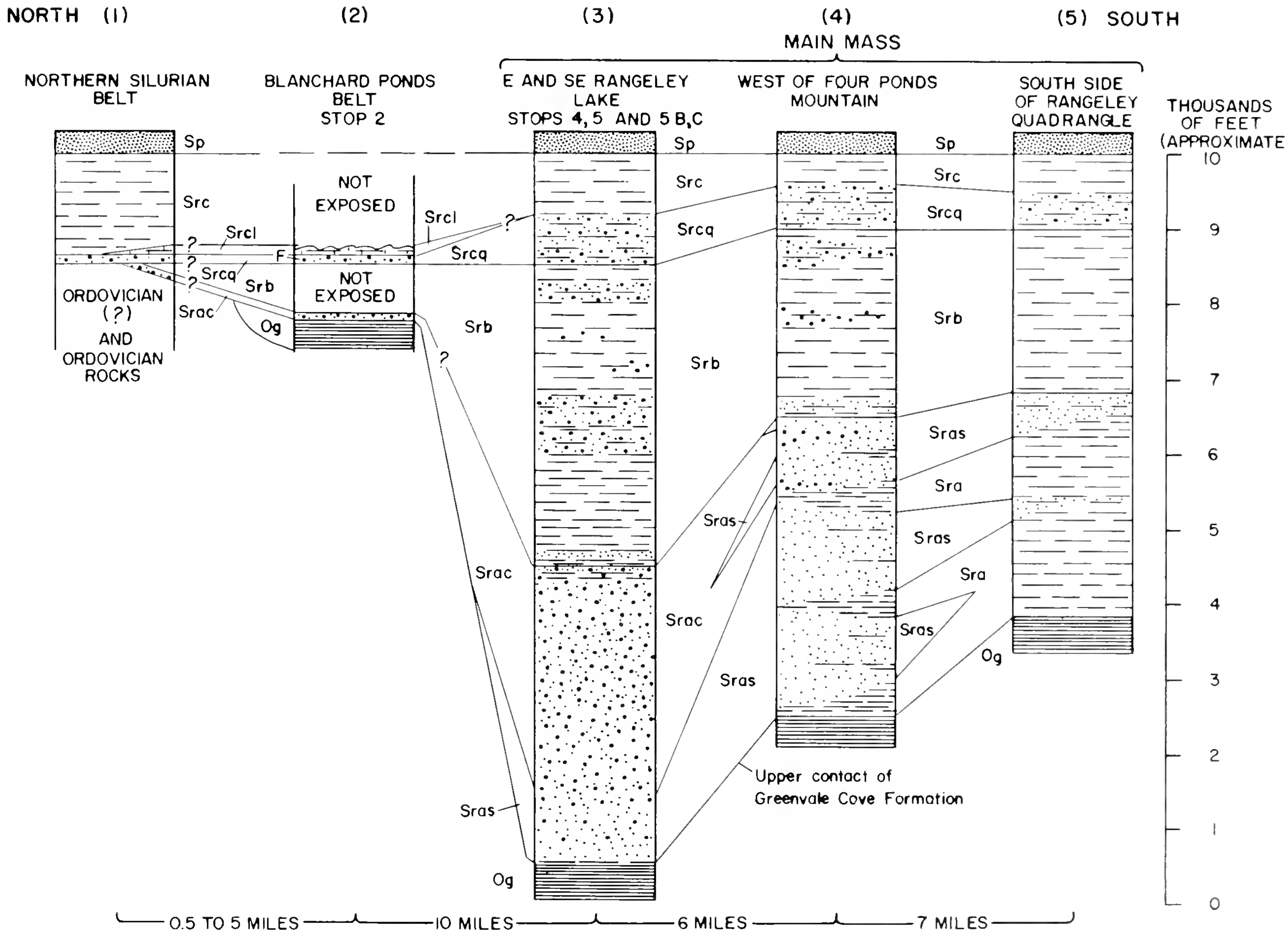


Figure 2.--Stratigraphic columns of the Rangeley Formation, Kennebago Lake (1, 2) and Rangeley (3, 4, 5) quadrangles. Unit symbols and descriptions on Figure 1. F indicates fossiliferous rocks at stop 2, Lower Silurian (upper Llandoverian, C_4-C_5).

Column 3 (fig. 2) includes the type localities of the Upper Ordovician(?) Greenvale Cove Formation and the Silurian(?) Rangeley Formation (Moench, 1969; Osberg and others, 1968, p. 251). Because this column was constructed from a thick, conformable, and well-exposed sequence of rocks, it serves as a measure for comparing other sequences to the north and south.

The Upper Ordovician(?) Greenvale Cove Formation is about 600 feet thick at the type locality and is 200 to 500 feet thick in the area of the Brimstone Mountain anticline. It is a distinctive unit composed of interlaminated, rather light-colored metasiltstone, and metasandstone. The silty and sandy rocks are commonly calcareous and form an extensive unit of calc-silicate rock in the area of the Brimstone Mountain anticline.

The Greenvale Cove is conformably overlain by 1,000 feet of coarse-grained feldspathic metasandstone grading upward to 3,000 feet of pebble, cobble, and boulder polymictic metaconglomerate--the Rangeley Conglomerate as used by Smith (1923), or the coarse facies of part A of the Rangeley Formation. These coarse clastics are characterized by typically massive beds as much as 30 feet thick, some of which channel deeply into underlying beds. Fragments are well rounded pebbles, and cobbles, a few boulders as much as 2 feet across, and sparse angular slabs of laminated metasedimentary rocks. They are set in a matrix of coarse grained metasandstone. Rock types are felsic and mafic metavolcanic rocks, vein quartz, quartzite, metamorphosed siltstone, sandstone, limestone (some of which is crinoidal), lamprophyre, and granitic rocks, including distinctive medium-grained granodiorite and quartz diorite with deformed dipyramids of blue quartz. Rocks that were metamorphosed prior to erosion and deposition have not been recognized. A few rounded clasts show evidence of predepositional weathering. Beds are crudely graded, commonly showing inverted grading in the lower foot or two, and normal grading from metaconglomerate to metasandstone in the upper several feet. Isolated pebbles and cobbles are widely scattered through the upper sandy parts of many graded beds. Most of each bed is massive but faint parallel lamination is present in metasandstone near the top of some graded beds. Crossbedding is uncommon. Clasts show no evidence of sedimentary imbrication. They are crudely alined subparallel schistosity, and elongated subparallel to bedding-schistosity intersections.

Part A is overlain by 4,000 feet (part B) of gray, commonly rusty-weathering metashale and graded interbeds of metasandstone in subordinate amount, and two conspicuous conglomeratic layers. Conglomerate clasts are similar to those of part A, but the proportions of stable quartzite and vein quartz to other rock types that are less stable in the zone of weathering is much greater than in the polymictic metaconglomerate of part A. The largest rounded clasts in the area represented by column 3 are cobbles and small boulders whose largest dimensions are generally less than 12 inches. Pebbles, cobbles and small boulders are commonly set in a matrix of pelitic

on semipelitic phyllite. Slump folds, intraformational unconformities, and conglomeratic mudflow deposits are characteristic.

Part C, about 1,500 feet thick, is incompletely exposed at stop 5B, C. Part C is dominately pelitic and similar to part B, except that the lower half of part C contains distinctive, commonly lenticular graded and locally crossbedded beds of quartz granule and pebble metaconglomerate, which locally contains abundant calc-silicate minerals (amphiboles, grossular, and clinozoisite). Conglomeratic rocks are typically better sorted than those of parts A and B of the Rangeley. Rounded fragments of vein quartz and quartzite as large as large pebbles are commonly closely packed; interstices are filled with coarse grained metasandstone and calc-silicate minerals. Knowledge of the sequence of part C is completed from outcrops west of Four Ponds Mountain (fig. 1), where part C grades upward into distinctive cyclically bedded cross-laminated quartzite and lightcolored muscovite-rich metashale of the Perry Mountain Formation.

Thick-bedded quartzite and quartz-granule conglomerates are present as well in parts of the Perry Mountain Formation and in the northern facies of the Smalls Falls Formation (fig. 1), but their sedimentary features and associations are unlike those of part C of the Rangeley.

South from the area of column 3 (fig.2), the Greenvale Cove Formation persists as a thin but distinctive unit, but the overlying Rangeley Formation changes dramatically. Polymictic conglomerates and feldspathic sandstones of part A become finer grained and less abundant, and they tongue southward into a thick mass of dominantly pelitic rocks. Conglomeratic rocks of part B likewise become finer grained and disappear southward. Quartz conglomerates of part C are more persistent, but also become finer grained and less abundant southward. In the area of column 5, the Rangeley Formation is, thus, a thick, rather monotonous mass of gray metashale interbedded at irregular intervals with subordinate amounts of metasandstone, and in part C with small amounts of quartz granule conglomerate. Farther south, the Rangeley is probably not divisible into parts A, B, and C.

At stop 2 (fig. 2, column 2), about 10 miles north of the area of column 3, rocks similar to those of the Greenvale Cove Formation are exposed less than 50 feet northwest of prominent outcrops of polymictic conglomerate, which is identical to the polymictic conglomerate of part A to the south. The largest fragments here are large cobbles, but boulders as much as 3 feet across have been found in other outcrops of polymictic conglomerate of part A in the Kennebago Lake quadrangle. Tops of bedding in both formations at stop 2 are interpreted to face southeast. Southeast-facing fossiliferous quartz conglomerate and interbedded baked shale and sandstone are exposed about 300 feet southeast of the polymictic conglomerate. The exposed rocks are comparable to those of part C of the Rangeley farther south (fig. 2, columns 2, 3, 4, 5). Shelly fossils, obtained from the localities shown on Figures 1 and 3, indicated a late Llandovery age for these quartz conglomerates. The fossil assemblage, as identified by

A. J. Boucot, includes:

Stricklandia lens ultima

Eocoelia hemispherica

Atrypa reticularis

Protomegastrophia(?)

Eocoelia cf. _E intermedia_

Farther to the southeast a thin unit of metamorphosed laminated argillaceous limestone abuts against intrusive rocks of the Stratton pluton (fig. 1). No comparable unit of metamorphosed impure limestone is present in part C farther south, but a few thin beds of metamorphosed fine-grained calcareous rock were found.

Despite obvious thickness changes in the Rangeley Formation, the exposed sections at stop 2 thus is comparable to the established Greenvale Cove and Rangeley sequence farther south (fig. 2, columns 2 and 3). The Greenvale Cove at stop 2 is much the same as it is farther south. Polymictic conglomerates at stop 2 thicken dramatically southward, become finer grained, and grade farther south to metamorphosed sandstone and then shale. The dominantly pelitic part B is not exposed at stop 2, but may be covered. It is interpreted to thicken southward (fig. 2). The fossiliferous quartz conglomerates and associated metasandstones and metashales at stop 2 may be equated with similar, though nonfossiliferous, rocks that form the lower half of part C farther south. The metamorphosed argillaceous limestone at stop 2 may be a rather restricted near-shore facies that did not extend far to the southeast. The upper half of part C is not exposed at stop 2.

Polymictic conglomerate that outlines the southern end of the northern belt of Silurian (?) rocks is tentatively correlated by us with the polymictic conglomerate at stop 2 (figs. 1, 2). At least two alternatives, however, are equally viable from present information: polymictic conglomerate of the northern belt is equivalent to (1) quartz-rich polymictic conglomerate in part B of the main mass; or to (2) quartz conglomerates of part C in the Blanchard Ponds belt and in the main mass. The second alternative is compatible with fossil data (Harwood and Berry, 1967, p. D-21), but would require a different source area for the various nonquartzose clasts in the polymictic conglomerate of the northern belt. An additional problem is involved in all three alternatives: Lower Silurian polymictic conglomerate in the Blanchard Ponds belt apparently conformably overlies the thin, but extensive, Upper Ordovician(?) Greenvale Cove Formation. In contrast, at the southwest end of the northern belt, steeply dipping upper Middle Ordovician black slates, which are stratigraphically well below the Greenvale Cove and Quimby Formations, are overlain by less deformed polymictic conglomerate; an angular

unconformity is inferred (Harwood and Berry, 1967, p. D-20). These contrasting belts of inferred conformity and unconformity are locally less than half a mile apart, (fig. 1). A sedimentary model that would account for these relationships is difficult to visualize, and conceivably the two belts have been faulted against one another. The northern belt, for example, could be an outlier of a large, nearly flat-lying thrust sheet. Available exposure, however, is inadequate to solve these problems.

Fossiliferous quartz conglomerate of the Blanchard Ponds belt is correlated with massively bedded vitreous-appearing quartzites and quartz conglomerates that outline the northwest side of the northern belt and are exposed on the east shore of Kennebago Lake, about 1 mile north of stop 1 (fig. 1). The quartz conglomerate on the east shore overlies a unit of dominantly pelitic rocks interbedded with two lenticular layers of polymictic conglomerate (fig. 3); this conglomerate is not unlike that of part B of the main mass to the south. Similar polymictic conglomerate is exposed as well in the wedge-shaped unit of part B(?) southwest of Kennebago Lake (fig. 1). Because good quartz conglomerates have not been found at the upper contact of the wedge-shaped unit, this contact is conjectural. This conjectural contact is within a mass of interbedded gray slate and feldspathic metasandstone which conformably overlies both the polymictic and the quartz conglomerates of the northern belt (Harwood and Berry, 1967, p. D-21) and which is similar lithologically to the dominant rocks of parts B and C in the Rangeley quadrangle. At the west edge of the northern belt, quartz conglomerate overlies polymictic conglomerate (Harwood and Berry, 1967, p. D-20), in accord with the established sequence to the southeast, but without the intervening pelitic rocks of part B. All these relationships, thus, tentatively suggest that parts A and B of the Rangeley are overlapped by part C toward the north (fig. 2).

Metamorphosed impure limestone at stop 2 is in turn equated with similar rocks that directly overlie quartz conglomerate to the north, near the southeast end of Kennebago Lake (fig. 1). Similar calcareous rocks are likewise associated with quartzite farther northeast in the northern belt.

Quartz conglomerate of the northern belt is overlain by at least 1,000 feet of irregularly interbedded gray metashale and feldspathic metasandstone. These rocks are characteristic of the upper part of part C of the Rangeley Formation to the south. In addition, at relatively high elevations northeast of Kennebago Lake (fig. 1), rocks correlated with part C of the Rangeley are locally overlain by cyclically interbedded pale-greenish-gray metashale and cross-laminated metasandstones that are remarkably similar to the light-colored Perry Mountain Formation of the Rangeley and Phillips quadrangles. These relationships strengthen the overall correlations that are illustrated on Figure 2.

The vitreous-appearing quartzite and quartz conglomerates of the northern belt typify the widespread upper Llandoverian Clough Quartzite of the Bronson Hill anticline to the southwest in New Hampshire (Billings, 1956; Boucot and Thompson, 1963). Moreover, the local presence of fine-grained

calcareous rocks above the quartzite is analogous to relationships between the calcareous Fitch and quartzitic Clough Formations in New Hampshire (Billings, 1956). Correlation of these units is thus extremely tempting, but should be avoided until age relationships among various exposures or belts of quartzite and calcareous rocks are more firmly established. Rocks assigned to the Clough Quartzite in New Hampshire, for example, may be of at least two different ages, according to Boucot (Pavlides and others, 1968, p. 65). The same might be true of the Fitch. In the area of trip A-1, for example, thick-bedded quartzites or quartz conglomerates are exposed in part C of the Rangeley Formation, in the middle and upper parts of the Perry Mountain, and in the northern facies of the Smalls Falls. Moreover, fine-grained calcareous rocks are conspicuous in the Greenvale Cove Formation, in part C of the Rangeley in the Kennebago Lake quadrangle, in the Madrid Formation, and in the Hildreths Formation of the southern sequence (fig. 1). Boudette favors correlation of the calcareous rocks of part C of the Rangeley in the Kennebago Lake quadrangle with the Fitch Formation, whereas Moench favors correlation of the Madrid Formation (which overlies the Smalls Falls and Perry Mountain, as well as part C) with the Fitch. If the Fitch Formation is, in fact, of more than one age, both authors may be correct.

Lower Silurian quartz conglomerates and some polymictic conglomerates are widely distributed in the northern Appalachians (Pavlides and others, 1968, table 5-1), but only in northeastern Newfoundland is there known to be a great thickness of polymictic plutonic-clast conglomerate comparable to part A of the Rangeley Formation. Marshall Kay (written commun., 1970) has called our attention to possible similarities between the Rangeley Formation and the Goldson Group, described by Helwig and Sarpi (1969), some 800 miles to the northeast. The provenance, composition and sedimentology of the Goldson Group is indeed remarkably similar to that of the Rangeley Formation. Both units are of approximately the same age. Although quartz conglomerates are lacking in the upper part of the Goldson, unit J of the Mix Cove Formation in the uppermost unit of the Goldson Group -- is composed of laminated, cross-laminated and convolute-laminated reddish and greenish siltstones (Kay, 1969, table 1; Helwig and Sarpi, 1969, p. 446-447) that are not unlike typical rocks of the Perry Mountain Formation. Environments in these distant areas may thus have been similar at approximately the same time.

Acknowledgments.--We are grateful to H. R. Dixon and R. W. Schnabel for constructive comments on the manuscript.

ROAD LOG FOR TRIP A-1

Assemble at the Rangeley Chamber of Commerce building well before the time of departure of your section. Owing to the added time required for stop 11, section A will leave at 6:00 a.m. Sections B and C will leave at 7:00 a.m., but will travel in opposite directions to stops 1 and 10, respectively. The trip will be by school buses. Owing to parking problems along state highway 4, and logistics at some stops, automobiles will not be permitted.

Participants in section A should be prepared for a 650-foot climb, 2-3/4-mile round trip to stop 11. The transition from part B into part C of the Rangeley Formation and then into the Perry Mountain Formation is will displayed here in extensive pavement outcrops. Participants in sections B and C should be prepared for a somewhat shorter (500-foot climb, 1-1/2-mile round trip) to stop 5B and C, where characteristic features of part C of the Rangeley are exposed in extensive pavements across the troughline of the Mountain Pond syncline.

Mileage

0.0 Start at corner of Pleasant and Main Streets (State Routes 16 and 4, respectively), opposite the Chamber of Commerce building, Rangeley.

Drive west on Main Street.

0.6 Turn right on Loon Lake Road and drive north.

1.2-1.4 Quimby Formation (Or) crops out on right side of road.

2.3 Pavement ends at Rangeley Airport on left.

2.4 Road intersection--drive straight ahead on right fork of gravelled roads.

2.8 Garnet-bearing reaction hornfels (included in Dga) in roof zone of Devonian gabbro crops out on right.

3.3 Caution: Sharp bend to right in road.

4.1 Ridge of Spotted Mountain (reaction hornfels) visible to northwest.

4.7 Borrow pit in spheroidally weathered gabbro (Dga) in pre-Pleistocene weathering zone.

5.3 Road junction: bear right from Loon Lake Road onto Kennebago Lake Club access road.

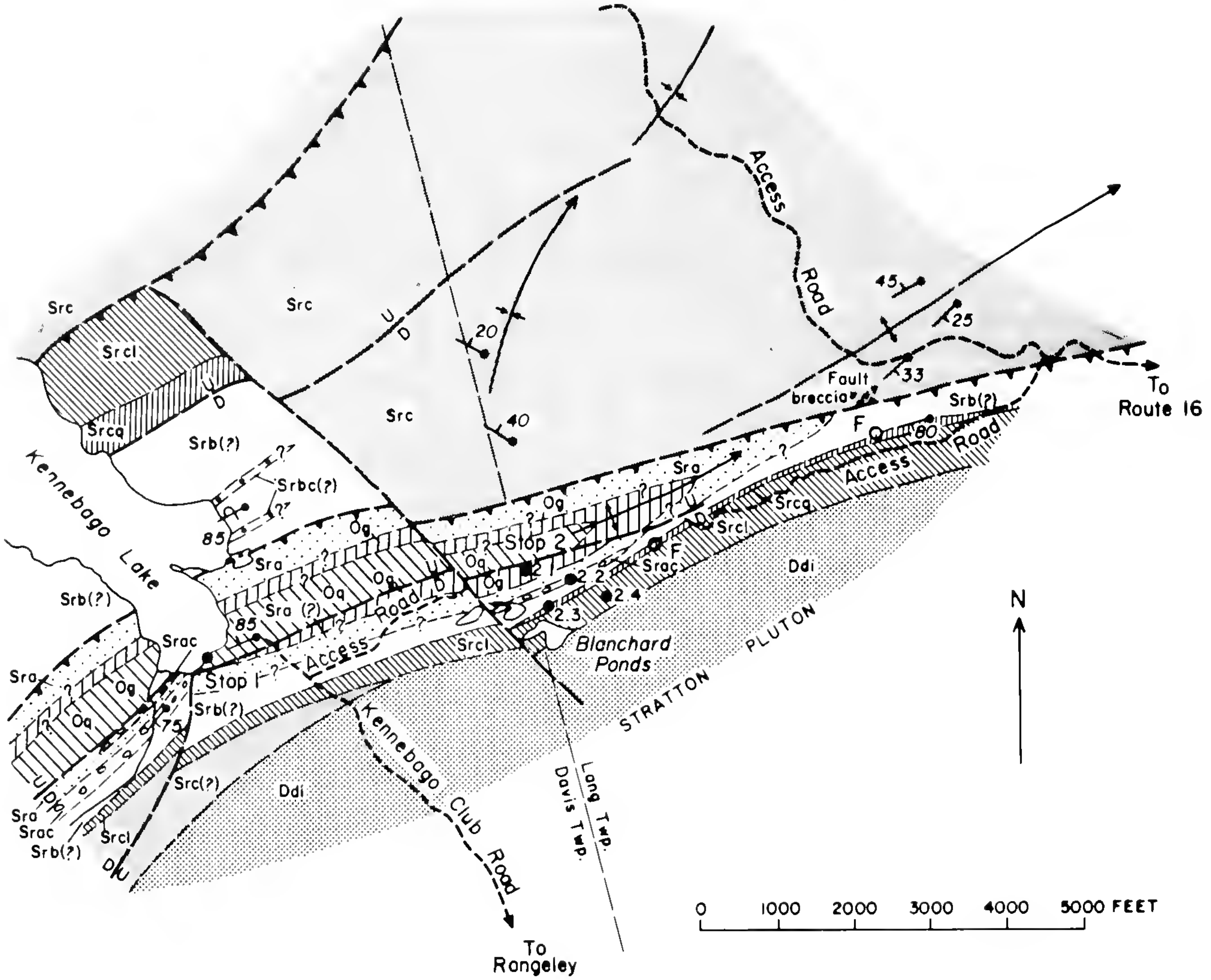


Figure 3.--Detailed geologic map of the Kennebago Lake area, Kennebago Lake quadrangle, showing location of stops 1 and 2. Planimetric base from unrectified aerial photograph (U.S. Geol. Survey Series GS-VBKX no. 1-66, 5/24/66, approx. scale 1:26,000). Geology by E.L. Boudette, assisted by D.S. Harwood, Jay Murray, and Woodrow Thompson, 1962-68.

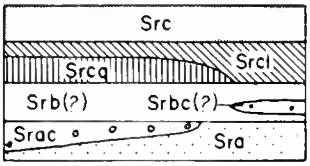
EXPLANATION

DEVONIAN

Medium- to coarse-grained hornblende diorite

Lower Silurian(?)

SILURIAN(?)



Rangeley Formation

Part C:

- Src, *pelitic andalusite-cordierite hornfels with thick- and thin-bedded arkosic metasandstone; may be part B* (Srb) *in part*
- Srcl, *thin-bedded and laminated calc-silicate rock and metamorphosed silty limestone*
- Srcq, *fossiliferous quartz pebble conglomerate and metasandstone east of Blanchard Ponds; vitreous-appearing, locally conglomeratic quartzite on north side of Kennebago Lake*

Part B:

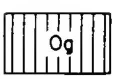
- Srb, *pelitic andalusite-cordierite hornfels with thick- and thin-bedded arkosic metasandstone; possibly same as part C* (Src)
- Srbc, *polymictic conglomerate; possibly same as part A* (Srac)

Part A:

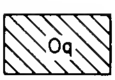
- Srac, *polymictic conglomerate*
- Sra, *thick- and thin-bedded metasandstone, possibly with some pelitic rocks*

Upper Ordovician(?)

ORDOVICIAN(?)



Greenvale Cove Formation



Quimby Formation

— — — - - ? - - -

Contact

Long dashed where approximately located; short dashed where inferred or gradational; queried where conjectural

U
D

Thrust

Teeth on upper plate

Normal

U, *upthrown side;*
D, *downthrown side*

Faults

Approximately located

Anticline

Syncline

Folds

Showing trace of axial surfaces and direction of plunge of axes. Approximately located

25

85

Upright

Overturned

Strike and dip of beds

Tops determined from sedimentary features

1

2.1

Location of outcrops at stops 1 and 2

F

Fossil locality

5.8 Garnet-bearing reaction hornfels in roof zone of Devonian gabbro crops out on right.

6.0 Borrow pit in till.

6.7 Outlet of Cow Pond (Note: Erratics along the road represent fairly well the underlying lithologies; they include reaction hornfels, gabbro, and diorite.)

10.4 Blanchard Ponds trail on right side of road.

10.6 Kennebago Lake Club--turn to left on service road and park in service area.

10.7 <u>STOP 1</u>. Outcrop on Lake shore to the west of the small stream which runs behind the service area.

Rocks exposed here are assigned to the upper part of the Quimby Formation, perhaps within 600 feet of the base of the Greenvale Cove Formation. This assignment is based on lithologic similarity to rocks in the type area of the Quimby (Moench, 1969) exposed at stop 3, and on the proximity of outcrops of the Greenvale Cove and Rangeley Formations at stop 2, about 4,000 ft east-northeast of stop 1, and of the Rangeley Formation about 800 ft southwest and 2,000 ft north of stop 1. (See fig. 3.) The dominant rock at stop 1 is dark, thin-bedded sulfidic slate, and lighter colored metagraywacke composed of quartz, plagioclase, subordinate amounts of biotite, and sparse magnetite and pyrite. The metagraywacke beds display fair to good graded bedding and poor cross laminations. Bedding strikes northeast, dips steeply, and faces northwest.

The rock at stop 1 is inferred to occur on the northwest limb of an isoclinal fold that trends about N. 70° E., and plunges at a low angle to the east (fig. 3). The Quimby on the southeast limb is interpreted to have been truncated by a normal strike fault along which the Rangeley Formation and perhaps a small amount of the Greenvale Cove have been downfaulted on the southeast against the Quimby. The Quimby Formation at stop 1 may be overlain to the north by the Greenvale Cove Formation; no exposures exist to confirm its presence. Farther north, along the east shore of Kennebago Lake (fig. 3) a succession is exposed, grading well up into the Rangeley Formation.

Rocks at stop 1 are within the contact metamorphic halo of the Stratton pluton. They contain biotite. More pelitic overlying rocks of the Rangeley and Greenvale Cove Formations exposed nearby contain andalusite and cordierite.

Return to buses.

10.7 Turn around, begin retracing rout back toward Rangeley, and park at woods access road on left 0.2 mile from Kennebago Lake Club.

0.2 STOP 2. Walk 0.7 mile to northeast along woods access road (fig. 3) to the limit of vehicle travel.

Borrow pits occur along road in valley train deposits probably associated with a large glacial lake in the Dead River and Sandy River drainages. Proceed about 1,200 feet, azimuth S. 81° E., from the end of the usable part of the road to an area of several scattered outcrops north of the largest Blanchard Pond (fig. 3).

A succession of four principal exposures occurs in apparently conformable normal stratigraphic order at stop 2 (fig. 3). In ascending stratigraphic order they represent: stop 2.1, the Greenvale Cove Formation; stop 2.2, polymictic Rangeley Conglomerate as used by Smith (1923) or the conglomeratic facies of part A of the Rangeley Formation; stop 2.3, fossiliferous quartz metaconglomerate and metasandstone correlated with the lower part of part C of the Rangeley (fig. 2); stop 2.4, calc-silicate rock, considered to be a local calcareous facies of part C of the Rangeley Formation (fig. 2). Dominantly pelitic rocks of part B of the Rangeley are not exposed, but may occur between stops 2.2 and 2.3. The calc-silicate rock is in contact on the southeast with mafic rocks of the Devonian Stratton pluton (fig. 1).

The Greenvale Cove Formation (Og) at stop 2.1 is exposed in small outcrops on the forest floor. Abundant angular rubble of Greenvale Cove is present nearby. The rock is laminated metasiltstone and fine- to medium-grained metasandstone, composed of quartz, plagioclase, and biotite. This rock is similar to the upper part of the Greenvale Cove at stop 4 (fig. 1).

Polymictic conglomerate (Srac) at stop 2.2 is exposed in an unusual series of 3-dimensional outcrops. Clasts are rounded, pebble-and cobble-sized fragments of granitic rocks, diorite, felsic and mafic volcanics, quartzite, quartz, and chert. They are set in a matrix of feldspathic metasandstone. Most of the clasts can be correlated with known pre-Silurian rock types exposed in the core of the Boundary Mountain anticlinorium to the northwest. The distinctive quartz diorite with blue quartz is exposed on Round Mountain in the Chain Lakes quadrangle to the north, and occurs in the metaconglomerate at stop 4 as well. Tabular clasts tend to be imbricate at stop 2.2. Most clasts, however, tend to be elongated parallel to the strike of bedding. Individual conglomerate beds

here are lenticular.

Fossiliferous rocks at stop 2.3 are thick bedded faintly laminated quartzose metasandstone and quartz-pebble metaconglomerate (Srcq). Individual beds commonly grade from quartz-pebble metaconglomerate at the base to metasandstone at the top. Shelly fossils, listed previously, occur in the conglomeratic, calcareous lower parts of the graded beds, and they are probably allochthonous. This fossiliferous unit is about 150 feet thick. It extends discontinuously along strike to the northeast almost to the eastern edge of the Kennebago Lake quadrangle, where it is cut by intrusive rocks. Efforts to trace it to the southwest have failed.

Calc-silicate rock (Srcl) exposed at stop 2.4 is believed to be metamorphosed silty to sandy limestone. The rock is distinctly laminated and is composed of various combinations of quartz, diopside, hornblende, garnet,epidote, chlorite, and small amounts of calcite. Mafic diorite of the Stratton pluton is exposed a few feet to the southeast.

Return to buses.

Return to Chamber of Commerce building in Rangeley.

0.0 Drive south along State route 4 from Chamber of Commerce building in Rangeley.

1.7 STOP 3. The shale member of the Quimby Formation (Or) exposed along Nile Brook. The rock is dark-gray pyrrhotite-bearing metashale (about 70 percent) with thin interbeds of metagraywacke and sparse thin beds of calc-silicate rock. Metagraywacke beds are graded, commonly parallel laminated, less commonly cross laminated; some have load casts. Tops of beds face northwest, on the northwest limb of an anticline whose axis is exposed on the east shore of Rangeley Lake.

Return to buses.

2.3 STOP 4. Starting at "The Terraces" cabins, walk south on left-hand side of road 1/2 mile past outcrops of Greenvale Cove Formation (Og) and part A of the Rangeley Formation (Sras and Srac).

First outcrops on the left are thinly interlaminated metashale and metasiltstone of the Greenvale Cove Formation (Og). Farther southeast, at a higher stratigraphic level in the Greenvale Cove, is a prominent zone of thinly laminated metasiltstone and metasandstone with abundant thin beds of calc-silicate rock. Metasiltstone and metasandstone laminations are commonly graded, but unequivocal tops are difficult to read in this lithology. Tops of rather sparse well-graded beds are interpreted to face southeast.

Basal feldspathic metasandstones of the Rangeley Formation (Sras) are exposed next to a telephone pole on the left. The basal 25 feet of the Rangeley is composed largely of coarse-grained light-colored metasandstone in graded beds facing southeast that range in thickness from a few inches to several feet. The metasandstone beds are separated by thin partings of gray metashale. The lower contact of the Rangeley is not exposed here in the roadside outcrops, but it is exposed on the hillside about 2,000 feet to the northeast. Features exposed there indicate that the basal Rangeley sands were deposited on unconsolidated shaly sediments of the Greenvale Cove Formation.

The lower 1,000 feet of the Rangeley--the sandstone facies of part A (Sras)--is abundantly exposed in the woods a few tens of feet east of Route 4. The dominant rock exposed here is massive feldspathic metasandstone in a succession of graded beds facing southeast. Single beds range in thickness from a few feet to 30 feet and tend to be thickest in the upper part of the sequence. Conglomeratic rocks are absent from the lower 250 feet of the sandstone facies. Small amounts of granule-pebble conglomerate first appear about 250 feet above the base and thence become gradually more abundant and coarser grained to the southeast toward the top of the sandstone facies.

Spectacular outcrops of conglomerate (Srac) begin opposite "The Birchwood" cabins and extend several hundreds of feet to the south along Route 4. The most impressive features of these rocks are the great thickness of individual beds (as much as 30 feet), their crudely graded, poorly sorted, and internally massive character, and the great variety of generally well rounded rock fragments. Unlike typical fluvial gravels, the largest clasts are not at the base of each bed, but instead are a foot or two above the base. Similar examples of inverted graded bedding have been noted by Fisher and Mattinson (1968) in turbidite-conglomerates at Wheeler Gorge, California.

Return to buses.

3.5 STOP 5. Turn left to Greenvale Cove School and park behind schoolhouse. Walk 100 yards northeast up gravel road to riprap quarry.

On northwest side of quarry are large blocks and outcrops of conglomerate, representing the lower of two prominent conglomeratic layers in part B of the Rangeley (Srb). The conglomerate is polymictic, but clasts of vein quartz and quartzite are much more abundant than in conglomerates of part A. In some rocks, interpreted as pebbly mudflow deposits, clasts are widely spaced in a matrix of pelitic phyllite.

Overlying rocks in the quarry and along the brook to the southeast are dominantly gray metashale and irregularly interbedded feldspathic metasandstone. Pebbly mudflow deposits, slump folds, and intraformational unconformities are exposed along the brook. Bedding tops face southeast, except on the northwest limbs of the

slump folds. Pelitic rocks contain euhedral pseudomorphs of chlorite and sericite after staurolite, which are locally alined subparallel to the plane of northwest-striking slip cleavage.

STOP 5B and 5C, for sections B and C only. Follow blazed trail (not shown on fig. 1) leading southeast from brook to pavement outcrops on hillside.

Part C of the Rangeley Formation (Srcq) is displayed in pavements that extend nearly 1,000 feet across strike. Rocks in the highest and farthest outcrops on the southeast are complexly folded in the axial zone of the Mountain Pond syncline. Fold axes plunge gently northeast to steeply southwest. Rocks in nearer and lower outcrops face southeast.

The dominant rock is gray metashale, which is irregularly interbedded with thin to very thick graded beds of feldspathic quartzite, and lenticular graded beds of closely packed quartz granule and pebble metaconglomerate. Beds of metaconglomerate range in thickness from a few inches to about 10 feet. Clasts are vein quartz and quartzite. Pebbly mudflow deposits are less common than in part B of the Rangeley.

Return to buses.

5.3 Long Pond on right.

6.0-6.1 Roadcuts of rusty-weathering dominantly pelitic rocks of part B of the Rangeley Formation (Srb).

6.4 Large road cut of rusty-weathering dominantly pelitic rocks. Dominant attitude of bedding is about N. 25° W., 25° NE, right side up, just southeast of the crestline of the Brimstone Mountain anticline. In contrast to the southwest plunge of the Mountain Pond syncline, the anticline plunges northeast.

6.7 South end of Long Pond.

7.1 Roadcut on right. Dominantly pelitic rocks of part B of Rangeley Formation (Srb).

9.2-9.5 Outcrops on right. Dominantly pelitic rocks of part C (Src). Rangeley Formation infolded with a small amount of the Perry Mountain Formation (Sp). Quartz pebble metaconglomerate of part C is exposed on the nose of an anticline that plunges steeply northeast.

9.8 Low outcrops on right. Uppermost part C of Rangeley Formation (Src).

10.6-10.8 Large outcrops on right. Perry Mountain Formation (Sp). Cyclically thin-bedded quartzite and light-colored muscovite-rich metashale and a few quartzite beds several feet thick. Bedding tops face northwest to the axis of a syncline at the extreme northwest end of the outcrops.

11.1 Bridge over Sandy River.

11.7-12.0 STOP 6. Watch traffic! Walk south on left-hand side of road past outcrops of uppermost Perry Mountain Formation (Sp) and lowermost rocks of Smalls Falls Formation (Ss). Tops of all beds face southeast.

Rocks of the Perry Mountain (Sp) are cyclically interbedded white quartzite and light-colored muscovite-rich metashale. Graded bedding is ubiquitous; cross lamination and convolute lamination are abundant.

Rocks of Smalls Falls Formation (Ss) are cyclically interbedded sulfide-rich rusty-weathering carbonaceous metashale and quartzite. Bedding features in lower part are similar to those of Perry Mountain.

Return to buses.

12.2-12.4 Outcrops on left; Smalls Falls Formation (Ss).

12.7 STOP 7. Smalls Falls picnic area; type locality of the Smalls Falls Formation (Ss). The formation is well exposed in cascades and pavements along the Sandy River above the falls and along the tributary Chandler Mill Stream a short distance to the west. Bedding features are best exposed along the Chandler Mill Stream.

The dominant rock type is rusty-weathering cyclically interbedded metashale and slightly subordinate quartzite. The metashale is dark-gray and carbonaceous and contains a few percent of pyrrhotite. The beds of quartzite, typically 1 inch thick, are well graded and commonly display a rather wispy cross lamination. The large altered and partly altered andalusite porphyroblasts tend to be alined subparallel to the plane of a pervasive slip cleavage that strikes northwest.

Return to buses.

13.0 Outcrop on left; Smalls Falls Formation (Ss).

13.4 Harvey Pond on left.

13.6-13.8 Large roadcuts on left. Smalls Falls Formation (Ss).

14.4 Bridge over Sandy River. Madrid Formation (Sm) exposed in stream bed; it is fine-grained light-colored metasandstone that contains abundant pods of calc-silicate rock.

15.2 STOP 8. Descend steep bank to outcrops along Sandy River. Contact between the Smalls Falls (Ss) and Madrid (Sm) Formations exposed on both sides of small island in center of stream. Tops of bedding face southeast.

Uppermost 500 feet of Smalls Falls (Ss, upstream) is a thin- to medium-bedded assemblage of sulfide-bearing calcareous metasandstones and metasiltstones, and a subordinate number of beds of noncalcareous metashale. Probably owing to their greater primary permeability, the coarser grained sediments of the original rock tend to be the most calcareous and contain various assemblages of calc-silicate minerals. The silty and shaly rocks are less calcareous, more carbonaceous, and darker colored. Some shaly beds contain as much as 75 percent almandine-spessartite garnet. Tops of beds near the upper contact of the Smalls Falls are indicated by graded bedding. The coarser clastics are lighter colored than the more carbonaceous finer clastics, but they are commonly softer than the finer clastics, owing to abundant calcite cement.

The contact between the Smalls Falls (Ss) and Madrid (Sm) Formations is sharp. It is marked by the abrupt appearance of coarser grained, lenticular-bedded, and crossbedded calcareous Madrid metasandstone. The abundance of sulfides diminishes gradually upward from the contact. The basal coarse clastic zone of the Madrid is about 35 feet thick. It is overlain by about 170 feet of interlaminated noncalcareous metasiltstone and metashale and some thin-bedded calc-silicate rock. The top of this shaly zone and younger rocks of the Madrid will be seen at stop 9.

Return to buses.

16.0 STOP 9, Madrid village. Type locality of the Madrid Formation (Sm) exposed along the Sandy River and the tributary Saddleback Stream. (See topographic map of Phillips quadrangle.). Walk south from the bridge over Saddleback Stream and along the east bank of the Sandy River.

The top of the shaly zone in the Madrid (described above in stop 8) is exposed north of the bridge over the Saddleback Stream. It is overlain to the southeast by 70 feet of thin-bedded light-gray, white, and pale-bluish-green calcareous metasiltstone and metasandstone. Note the distinctive cross lamination in the white beds. These rocks are overlain by 5 to 10 feet of sulfidic carbonaceous metasiltstone and metashale, followed by about 550 feet of thick-bedded light-gray to light-brownish-gray calcareous metasandstone with subordinate partings of medium-gray metashale. The beds of metasandstone are as much as 10 feet thick; they are graded, commonly crossbedded, and they commonly contain pods of calc-silicate rock. Gray metashale is most abundant in the southernmost outcrops along the Sandy River where the Madrid (Sm) grades

upward into the Seboomook(?) Formation (Ds). All beds face southeast.

Return to buses.

17.0 STOP 10. Seboomook(?) Formation (Ds) is exposed along the Sandy River under the bridge and downstream.

The rock is dominantly gray faintly bedded to well bedded metashale with greatly subordinate amounts of light-colored metasiltstone and fine-grained metasandstone. Tops of beds face northwest toward the trough of the Bear Hill syncline (fig. 1). Typical graded bedding is expressed by continuous gradations from sandy or silty rock into metashale. It is unlike graded bedding in the Rangeley or Perry Mountain Formation, for example, in which the contact between each graded bed of metasandstone and the next overlying bed of metashale tends to be sharp or sharply gradational. Where silty or sandy material is absent from the Seboomook(?), graded bedding is expressed by subtle gradations in staurolite content, which reflect changes of original alumina (or clay) content within each graded bed.

Return to buses.

STOP 11. Park near bridge over Sandy River (elevation 1,447 ft; see topographic map, Rangeley quadrangle). Extensive pavement outcrops are 5,000-6,000 feet, azimuth N. 85° W., from the bridge and at an elevation of about 2,100 feet. Their approximate location is shown on the geologic quadrangle map (Moench, 1970). Figure 4 illustrates stratigraphic and structural relationships of the upper part of the Rangeley Formation and the lower part of the Perry Mountain Formation.

The most important stratigraphic features at stop 11 are the distinctive beds of quartz metaconglomerate in part C of the Rangeley, and the gradational contact between the Rangeley Formation and the overlying Perry Mountain Formation. The quartz metaconglomerates are equated approximately with the fossiliferous quartz conglomerate at stop 2 in the Kennebago Lake quadrangle. At stop 11 (fig. 4) quartz metaconglomerates crop out at three stratigraphic levels and are separated by typical gray Rangeley metashales. The metashales are irregularly interbedded with generally greatly subordinate amounts of metasandstone. The metaconglomerates, together with associated coarse-grained metasandstone, form thick, commonly lenticular graded beds. Some of these coarse clastic rocks contain abundant clinozoisite, grossular, and amphibole, suggesting that the original cement was a mixture of carbonate and clay minerals. Pebbly mudflow deposits, composed of conglomerate clasts rather widely spaced through a matrix of metashale, are present but not abundant.

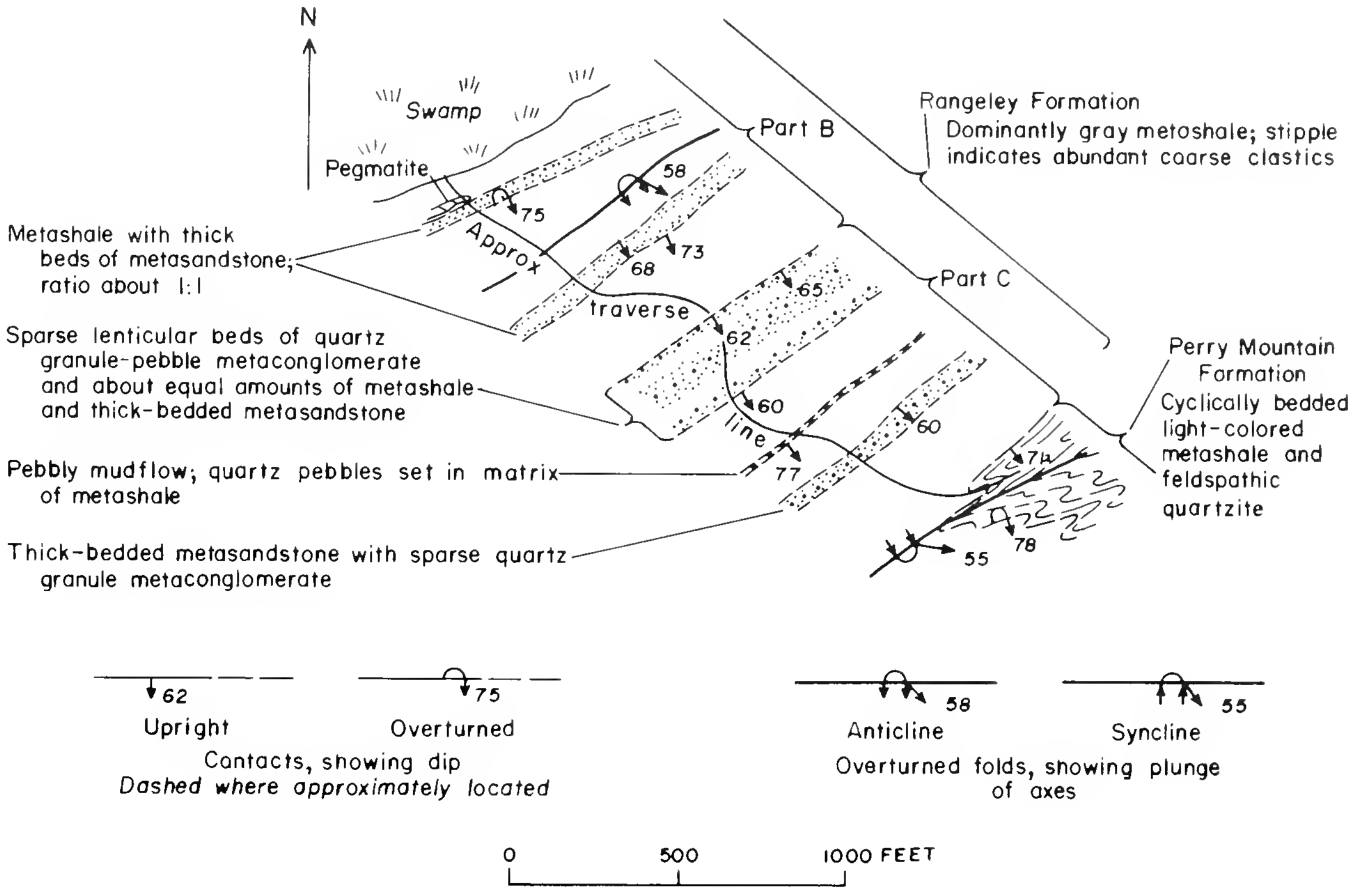


Figure 4.--Pace and compass sketch map of pavement outcrop at stop 11.

Gray Rangeley metashale that is rather irregularly interbedded with typically massively graded beds of feldspathic metasandstone grade upward into light-colored muscovite-rich Perry Mountain metashale that is cyclically interbedded with distinctively cross laminated, graded beds of more quartzose metasandstone.

The overturned folds shown on Figure 4 plunge steeply, locally directly down the dip of the phyllitic axial surface cleavage. The pelitic rocks are phyllites with porphyroblasts of staurolite, altered to chlorite and white mica.

REFERENCES CITED

Billings, M. P., 1956, Bedrock geology, pt. 2 of The geology of New Hampshire: Concord, N. H., New Hampshire State Planning and Development Commission, 203 p.

Boucot, A. J., and Thompson, J. B., Jr., 1963, Metamorphosed Silurian brachiopods from New Hampshire: Geol. Soc. America Bull., v. 74, no. 11, p. 1313-1334.

Fisher, R. V., and Mattinson, J. M., 1968, Wheeler Gorge turbidite-conglomerate series, California--Inverse grading: Jour. Sed. Petrology, v. 38, no. 4, p. 1013-1023.

Harwood, D. S., 1966, Geology of the Cupsuptic quadrangle, Maine: U.S. Geol. Survey open-file rept., 259 p.

Harwood, D. S., and Berry, W. B. N., 1967, Fossiliferous Lower Paleozoic rocks in the Cupsuptic quadrangle, west-central Maine, in Geological Survey Research, 1967: U.S. Geol. Survey Prof. Paper 575-D, p. D16-D23.

Helwig, J., and Sarpi, E., 1969, Plutonic-pebble conglomerates, New World Island, Newfoundland, and history of eugeosynclines, in Marshall Kay, ed., North Atlantic-Geology and Continental Drift: Am. Assoc. Petroleum Geologists Mem. 12, p. 443-466.

Kay, Marshall, 1967, Stratigraphy and structure of northeastern Newfoundland bearing on drift in North Atlantic: Am. Assoc. Petroleum Geologists Bull., v. 51, p. 579-600.

______1969, Silurian of northeast Newfoundland coast, in Marshall Kay, ed., North Atlantic - Geology and Continental Drift: Am. Assoc. Petroleum Geologists Mem. 12, p. 414-424.

Moench, R. H., 1969, The Quimby and Greenvale Cove Formations in western Maine: U.S. Geol. Survey Bull. 1274-L, p. L1-L17.

______1970, Geologic map of the Rangeley and Phillips quadrangles, Franklin and Oxford Counties, Maine: U.S. Geol. Survey Misc. Geol. Inv. Map I-605 (in press).

Osberg, P. H., Moench, R. H., and Warner, Jeffrey, 1968, Stratigraphy of the Merrimack synclinorium in western Maine, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., Jr., eds., Studies of Appalachian geology, northern and maritime: New York, Interscience, p. 241-253.

Pavlides, Louis, Boucot, A. J., and Skidmore, W. B., 1968, Stratigraphic evidence for the TAconic orogeny in the Northern Appalachians, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., Jr., eds., Studies of Appalachian geology--Northern and Maritime: New York, Interscience, p. 61-82.

Smith, Edward S. C., 1923, The Rangeley conglomerate [Maine]: Am. Jour. Sci., 5th ser., v. 5, p. 147-154.

U.S. Geological Survey, 1965, Geological Survey Research 1965: U.S. Geol. Survey Prof. Paper 525-A, 376 p.

STRUCTURAL RELATIONS IN THE KINGFIELD QUADRANGLE, MAINE

by

John Raabe
Northwestern Connecticut Community College

An investigation of the northern Kingfield quadrangle, Maine was initiated with support from the Maine Geological Survey in June 1968. Mapping by Robert Moench in the Rangeley and Phillips quadrangles and by Gary Boone in the Little Bigelow Mountain quadrangle prompted the investigation. Moench (1968) had delineated a large area of cyclically bedded pelitic sandstone, the Perry Mountain Formation of probable upper Silurian age (Osberg, Moench, and Warner, in Zen, 1968), on strike with Seboomook (?) rocks mapped by Boone (report in preparation) in the Carrabassett River Valley of the Little Bigelow Mountain quadrangle. As shown in fig. 1, an irregular southward projection of the Sugarloaf Gabbro (GA) separates the two formations by more than a mile in the northeast strike direction.

A well defined unconformity of Taconic age has been mapped west of the Rangeley-Phillips area (reviewed by Pavlides and others, 1968). In the Rangeley, Phillips and Kingfield quadrangles, however, no unconformity has been found, and in its place occurs a thick conformable sedimentation according to Osberg and others (1968). The Seboomook (?) rocks, here deposited upon a thick sedimentary pile, were locally raised by Early to Middle Devonian tectonism. Nevertheless downwarping, presumably related to Devonian plutonic emplacement, has preserved a veneer of Seboomook (?) rocks in the strike belt crossing the valley of the Carrabassett River north of Kingfield village.

At the outset, detailed structural mapping in the NW ninth of the Kingfield quadrangle was hindered by the general lack of dependable primary features. Shearing dominates the fabric of the pelitic rocks and had either obliterated classical top criteria or at least had rendered them speculative. As the investigation continued, a belt of severe slipping and shear was delineated. The belt is characterized by zones of 1-1 1/2 inch-wide pinstriped fabric, but appears massive where shearing is severe. Earlier premetamorphic soft sediment slumping described by Moench (1970) may have mixed the sediments, destroying any trace of bedding before the later shearing event. In many outcrops, short veins of quartz, 2 - 4 feet in length, are deformed with symmetric relationship to the slip cleavage. Excellent examples of the pinstriped fabric with associated quartz veins can be seen at the roadside along Deer Farm Road and at the Tufts Pond gaging station in the northwestern corner of the Kingfield quadrangle. Impressive examples of shear folding with the graded bedding preserved can be seen in pavement outcrop along the southwest flank of Vose Mountain on a small knob locally called Mossmann's Hill.

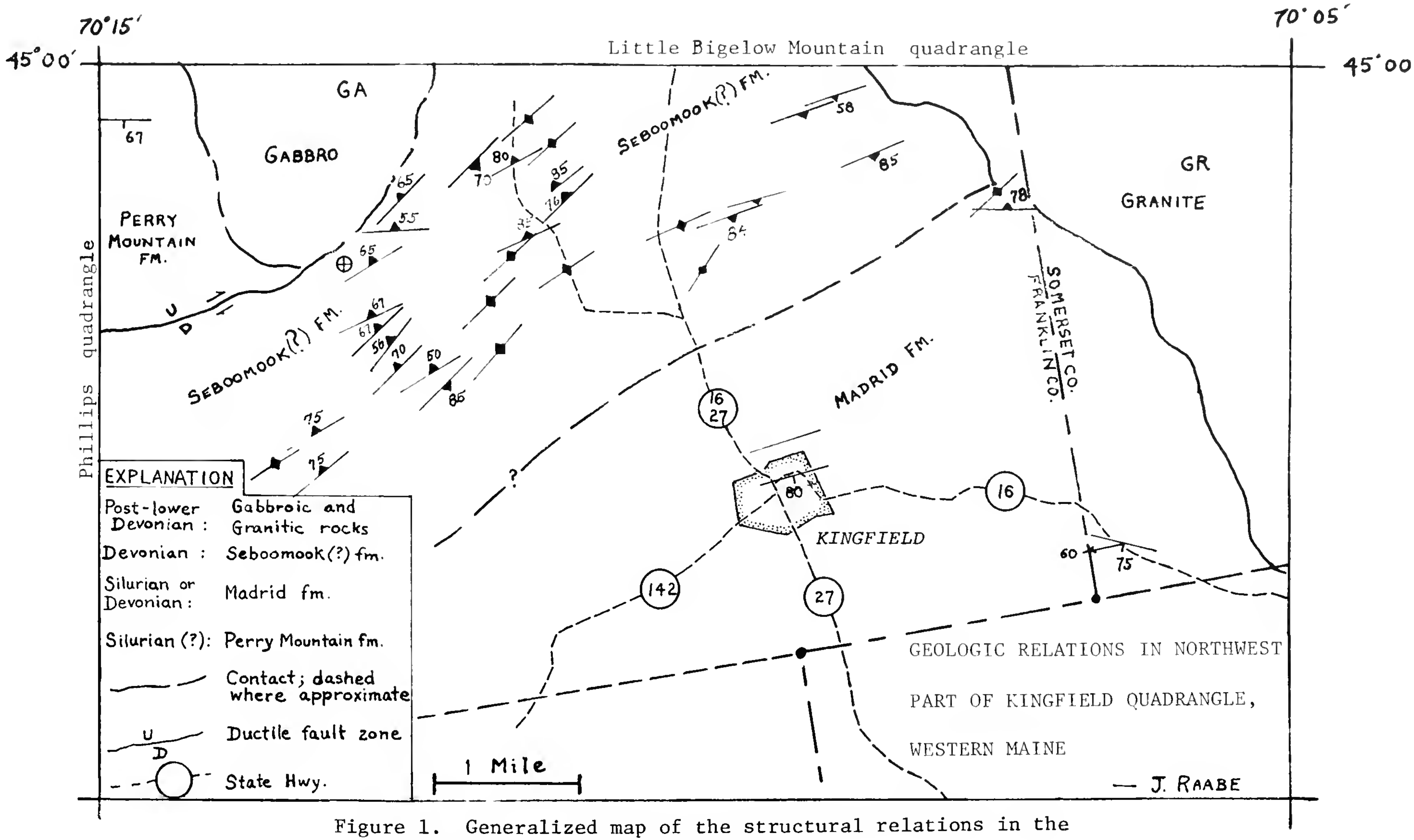


Figure 1. Generalized map of the structural relations in the northwestern part of the Kingfield quadrangle, Maine. The ⊕ indicates the location for fig. 2.

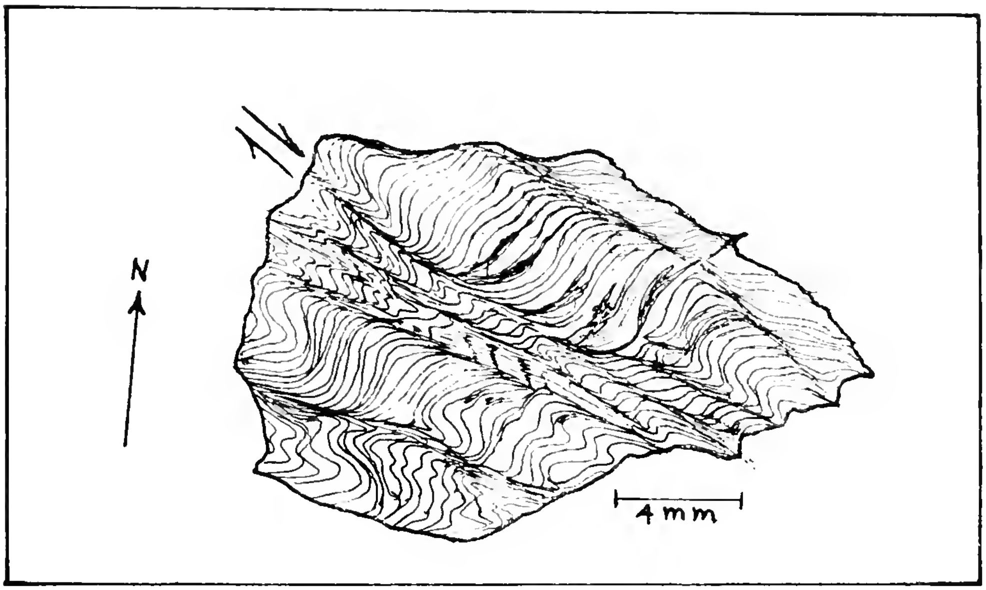
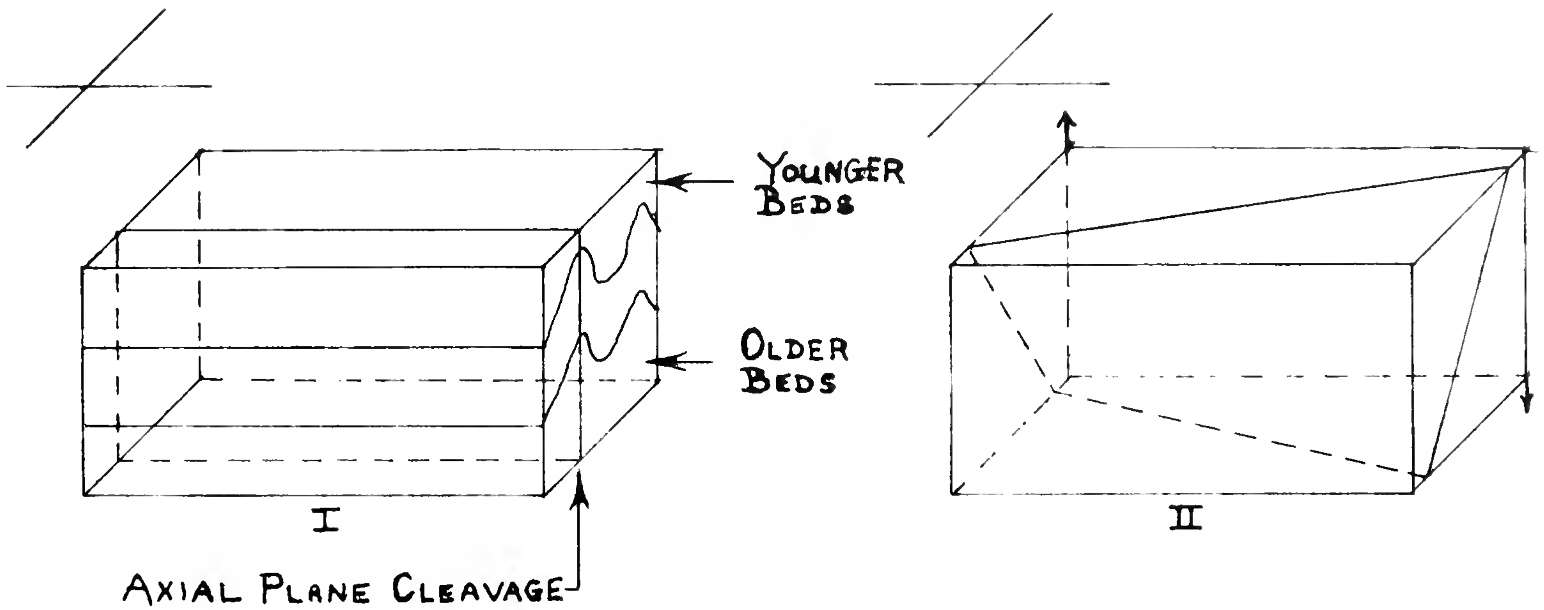
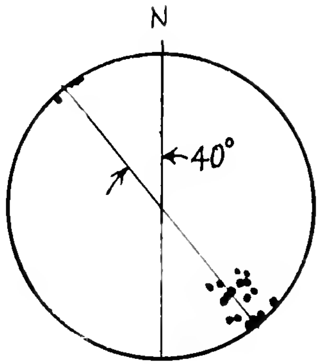


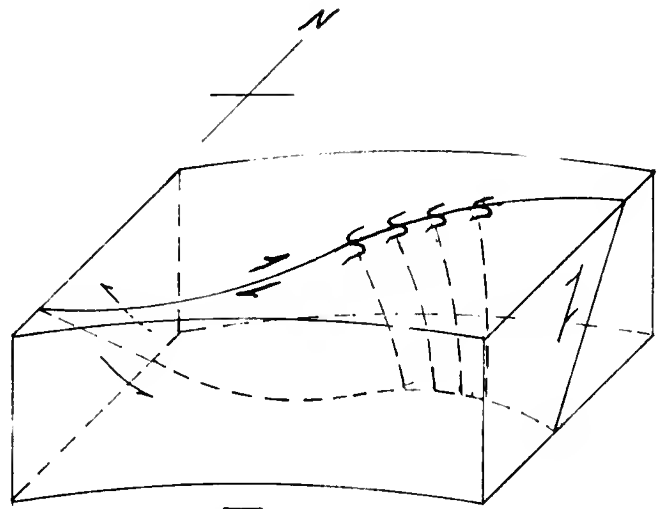
Figure 2. Sketch of typical rock fabric in the Seboomook (?) formation south of gabbro contact showing the earlier slaty cleavage at approximately N 60°E intersected by slip cleavage at approximately N 70°W. Elsewhere, both cleavages generally coincide at approximately N 45-60°E.

Figure 3. Model showing some of the structural relations in the northwestern part of the Kingfield quadrangle, Maine.





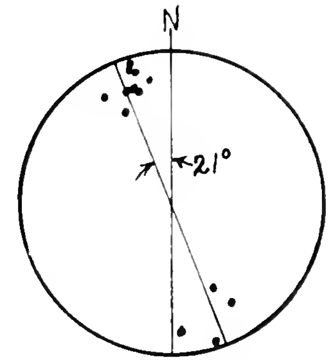
LOWER HEMISPHERE PLOT OF POLES TO FOLIATIONS IN VICINITY OF TUFTS POND BROOK. (WEST)



III

DIAGRAMMATIC REPRESENTATION OF THE DUCTILE SLIP ZONE IN N-W KINGFIELD QUADRANGLE, MAINE.

REPRESENTS ORIENTATION OF LINEATION IN THE EAST — NOT A CRENULATION



LOWER HEMISPHERE PLOT OF POLES TO FOLIATIONS IN AREA OF CARRABASSETT VALLEY. (EAST)

Along much of the shear belt, the shearing reactivated movement along earlier planes of weakness much as described by Ramsay (1967). On the scale of the outcrop, slaty cleavage served as the weakness while the regional extent of the belt is presumed to have been determined by previous tectonic control. The Barnjum Fault, a premetamorphic fault described by Moench (1970), has been extended to the southern tip of the gabbro as shown in fig. 1 by Raabe (1969). The premetamorphic slump may serve as the control on regional location of the slip zone. The role of the gabbro in the development of the zone is uncertain, but the geometric fit of its extension with the curvature of the belt is felt to indicate kinship.

Fortunately, in the vicinity of Tufts Pond Brook, the slip cleavage is not simply a reactivation of the earlier slaty cleavage but locally intersects it as shown in fig. 2. The intersection produces crenulations which show a clockwise rotational sense implying dextral movement. Maximum displacement took place normal to the plunge of the crenulations. Since the crenulations plunge moderately to the northeast, older rocks outcrop to the northwest.

In the area of the Carrabassett River valley, the movement has been in the plane of steeply dipping foliation and bedding. The resultant lineation, in which certain porphyroblasts are aligned, plunges moderately to the southwest (see Boone, Trip A2, this volume). Neither the lineation nor the slickensides with which it commonly coincides shows any persistent sense of movement. A condition of general collapse apparently prevailed in the area with attendant faulting, shearing, and bizarre folding. Only the near-vertical limbs of tight folds are commonly preserved and broad hinges are "sheared out".

Figure 3 is a model to demonstrate the relationships observed in the NW 1/9 and NC 1/9 of the Kingfield Quadrangle. In frame I, the "younger beds" represent the Seboomook (?) rocks, "older beds" represent the pre-Taconic rocks, and the unlabeled portion represents the metasedimentary wedge which replaces the Taconic break locally. The folding shown in frame I produced the slaty cleavage represented in figure 2. Rotation and flexure are shown in frames II and III respectively; however, both were probably produced in a single tectonic event associated with the emplacement of the post Lower Devonian plutons.

REFERENCES

Moench, R.H., (1968) Geologic Map of the Rangeley and Phillips Quadrangles: Open-file Report, U.S. Geol. Survey.

Moench, R.H., (1970) Premetamorphic Down-to-Basin Faulting, Folding, and Tectonic Dewatering, Rangeley Area, Western Maine: Geol. Soc. Amer. Bull. 81:1463-1496

Osberg, P.H., R.H. Moench, and Jeffrey Warner (1968), Stratigraphy of the Merrimack Synclinorium in West-Central Maine: p. 241-253 in Zen, E-an and others (1968).

Pavlides, Louis, A.J. Bucot, and W.B. Skidmore (1968), Stratigraphic Evidence for the Taconic Orogeny in the Northern Appalachians: p. 61-82 in Zen, E-an and others (1968).

Raabe, John (1969) Structural Relationships in the Northwestern Kingfield Quadrangle, Maine: Unpublished Masters Paper, Boston University.

Ramsay, J.G. (1967) Folding and Fracturing of Rocks, McGraw-Hill Book Co., New York, p. 546-549.

Zen, E-an, W.S. White, J.B. Hadley, and J.B. Thompson, Jr. (eds.) (1968) Studies in Appalachian Geology - Northern and Maritime: John Wiley and Sons, New York 475 p.

Tectonic and Metamorphic History of Lower Devonian Rocks in the Carrabassett Valley North of Kingfield, Maine

Gary M. Boone
Department of Geology
Syracuse University

Introduction

The polymetamorphic, complexly folded rocks of this part of the Carrabassett valley are instructive from the points of view of linking stratigraphies that are interrupted along strike by large plutonic bodies, and of witnessing a regional development of contact metamorphic assemblages produced in two recognizable dynamothermal events. On the one hand, stratigraphic and structural observations gathered in this and adjacent areas flanking batholithic plutons have each reinforced the other to provide a stratigraphic synthesis which can now be used to relate some of the Siluro-Devonian formations in the Merrimack synclinorium across a paleotectonic hinge to formations of similar age in the Moose River synclinorium (see introductory article, this guidebook). On the other hand, the areal distribution and textural relationships of metamorphic mineral assemblages in the valley tell much about the concurrence of thermal events with the crowding and rearrangement of simpler regional structures during the forcible emplacement of the Sugarloaf gabbroic massif to the west, and the Lexington granitic batholith to the east.

The setting for this trip is thus one of a complexly folded metasedimentary sequence straddling the north-south trend of the Carrabassett River. The terrain in which this sequence is exposed is referred to here as the "Carrabassett corridor". The igneous rocks that form the structural walls of the corridor can be considered as the opposing jaws of a tectonic vice on which can be blamed most of the kinematic happenings preserved in the rocks to be witnessed during the trip.

Acknowledgments

The Maine Geological Survey supported the field mapping. I am indebted to Robert G. Doyle for his continued support, and to geologists of the U.S. Geological Survey, particularly E.L. Boudette, R.H. Moench, and A. Griscom for discussions in the early stages of mapping that allowed me to benefit from their work already in progress in adjoining quadrangles. A study of late-stage retrogressive metamorphism and metasomatism in the rocks at stop 6 was carried out by

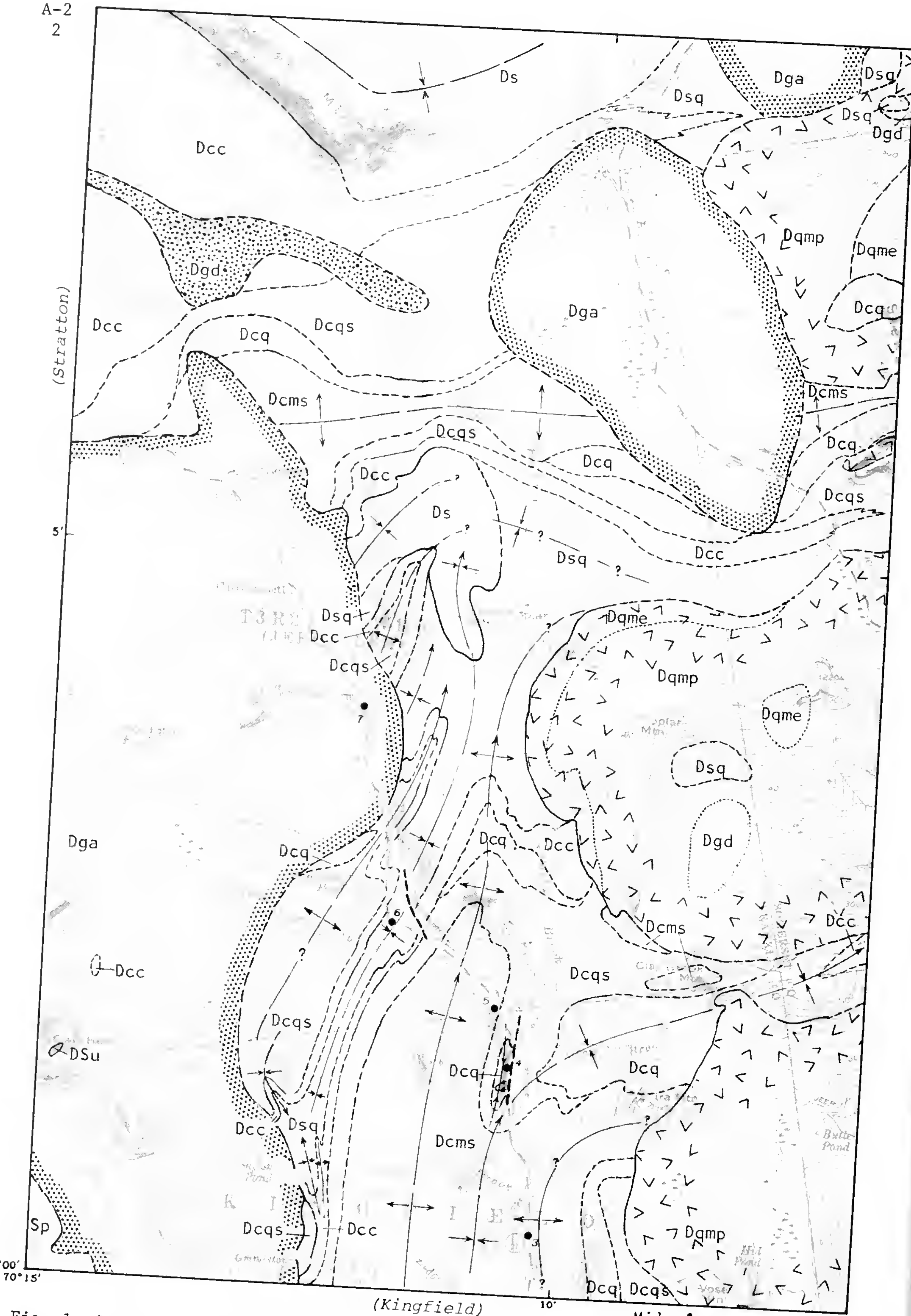


Fig. 1. Structure and stratigraphy of Carrabassett valley region, Little Bigelow Mountain quadrangle

Figure 1 EXPLANATION

Igneous Rocks

Post-lower Devonian

Lexington Batholith

- Dqme c.g. equigranular quartz monzonite
- Dqmp q.m. porphyry (roof and wall facies)
- Dgd granodiorite and diorite (ornament for batholith margins only)

Huston Flats granite-diorite stock

Late Acadian

Sugarloaf gabbroic massif

- medium- to coarse-grained gabbro, gabbro pegmatite, norite, and gabbroic anorthosite.
- Layered in part (ornament for massif margins only)

[Dikes and sills of gabbro, diabase, and garnetiferous aplite not shown]

Structural & Logistic Symbols

Contact
Dashed where inferred or concealed;
dotted where gradational

Inferred fault

Anticline, showing direction of plunge
Queried where inferred

Syncline
normal and overturned

5 ● Outcrop areas described in road log for trip A-2

Metasedimentary Rocks

lower Devonian

Ds — Seboomook formation

Upper member: cyclically graded-bedded metagraywacke and metashale (andalusite-cordierite schist or retrograde equivalents and sillimanite granofels in Carrabassett valley and Bigelow Range). Graded sets 5 cm to a meter thick; 10-15 cm common. Pelitic portions gray to dark gray, graywacke portions gray to light gray, on fresh surfaces.

Dsq — Schist and graywacke lower member: prominantly graded bedded in Carrabassett valley; lenticular and sheared within 2500 meters of Lexington batholith contact. Staurolite + garnet (chiefly as relicts) at, and SW of Rt. 16 - 27. Aluminous layers andalusite-bearing throughout most of area. Graded sets 5 to 30 cm thick.

lower Devonian (?)

Dcc — Carrabassett Formation (informal designation)

Sulfidic calc-silicate and schist member; proportion of calc-silicate to pelitic schist varies widely. Consistently sulfide-bearing. Calc-silicate rocks thin- to medium- layered. Associated non-porphyroblastic muscovite-biotite-plagioclase schist. Minor thin quartzite layers.

Dcq — Biotitic metaquartzite member. Variously feldspathic; thickly bedded or massive. Minor thick zones of massive gray schist. Metaquartzite dark gray where fresh; weathers to medium gray.

Stratigraphic position varies from uppermost part of Dcms to Dcc. Forms thick, laterally discontinuous lenses.

Dcqs — Thinly layered andalusite-cordierite schist and non-porphyroblastic quartz-rich schist. Rare staurolite relicts in sub-porphyroblastic schist layers. Poorly developed graded bedding.

Dcms — 'Massive' andalusite schist member; relict bedding very indistinct, or lacking; quartz-rich segragation-laminae (paper-thin to 2 mm thick, 3 to 6 cm long) developed along axial plane cleavage. Sulfide (as sulfide-quartz lenticles) very rare. Restricted sections, up to 50 feet thick, of graded-bedded metapelite and metagraywacke.

Silurian or Devonian (?)

DSu — Cordierite-rich, hypersthene-spinel rafts or roof pendants in Sugarloaf massif.

Silurian (?)

Sp — Perry Mountain (?) formation

Thin- to medium- graded-bedded muscovite-rich sillimanite-almandite granofels.

P.T. Lyttle, then an undergraduate at Syracuse. To the residents and land owners of the Carrabassett valley I express my thanks for their cooperation.

"Margaret and Frike"

As man becomes aware of his perpetrations upon nature, probably farthest from mind is the despoiling of rock outcrops with paint, let alone hammering. Some of the most spectacular outcrops, geologically and scenically, in the Carrabassett valley are extensively marred from the use of pressurized paint spray-cans. The 'half-life' of sprayed enamel is not yet measurable after four years' exposure. Despite this condition of some of the outcrops, I ask participants to refrain from hammering the most instructive (often weathered) surfaces.

Stratigraphy and Structure

The stratigraphic and structural synthesis in the Carrabassett corridor, as in many metamorphic terrains, has consisted of a cyclic integration of local structural geometries and lithologic successions. Each has led to a clearer understanding of the other, often from widely separated parts of the area. The relative inaccessibility of many key outcrops with regard to the time available for this trip has of necessity been the deciding factor in choosing selected outcrop illustrations of the principal problems, and there to share with participants the techniques and reasoning employed in the attempt to solve them. With such prospects of minor incisions to expose the anatomy of the corridor, its general structure and evolutionary history are herewith 'unfolded'.

Upon entering the Carrabassett valley at Kingfield, the structure of the Siluro-Devonian units deviates little from the regional northeasterly trend (see Raabe, this volume). Stratigraphic relations in rocks to the north in the Carrabassett corridor (Little Bigelow Mountain quadrangle), are based on the continuity of the section here and to the east of the Lexington batholith (fig. 1).

The plunge of major fold axes (not lineations) from the southern quadrangle boundary to Little Poplar Mountain is generally northeasterly. Together with top-facing depositional criteria in rocks near unit boundaries, these observations lead to the interpretation of progressively younger strata encountered in following the plunge of folds north to the vicinity of Little Poplar Mountain. These strata comprise units which are not recognized, or not separately defined as units, to the south. Brief mention of the succession is necessary here.

At the northern boundary of the Madrid Formation local faulting has disrupted the contact with overlying pelite. In eastern Little Bigelow Mountain quadrangle, however, the contact is gradational by way of interbedding across ~ 100 feet of section, and top-facing criteria (cross-bedding in the Madrid, and graded-bedding in the

overlying pelitic sequence) have been recognized very close to the contact where intervening minor folds are believed absent.

The immediately overlying pelite and associated local units in the Carrabassett valley are to be formally designated in a forthcoming publication as the Carrabassett Formation; it consists, in ascending order, of "massive" pelite, thinly interbedded pelite and quartz-rich pelite, thick, discontinuous lenses of massive or thickly bedded quartzite and minor thick zones of pelite, topped by thinly bedded, characteristically sulfidic aluminous calc-silicate and pelite which is quite variable in these lithic proportions along and across strike.

The massive quartzite member appears to be similar, in stratigraphic position and in sedimentary disposition to surrounding units, to the Whisky quartzite of the Moose River synclinorium as described by Boucot (1969).

The highest stratigraphic unit of the valley (and indeed, of the entire Little Bigelow Mountain quadrangle) into which the Carrabassett Formation grades abruptly, is composed of graded bedded metagraywacke and metapelitic rock, and quartzite. This unit is correlated with the Seboomook Formation of the Moose River synclinorium as defined by Boucot (1969). The proportion of quartzite and metagraywacke to metapelite decreases from the lower part of the formation to its highest exposed part. We will see the lower part of the formation on this trip.

The distinctions made between the pelitic parts of the Carrabassett and Seboomook (?) formations is not as clear-cut in detail as portrayed in this brief summary. Alternations in bedding characteristics commom to both formations are present in each, and it is my subjective integration of observed proportions in each unit that is used to separate them. Stratigraphic "lumpers" and "splitters" can square off on nearly every outcrop of the itinerary.

Relict features of turbidite deposition abound in the graded sets of the upper part of the Carrabassett, and in the Seboomook (?) Formation. Convolute "type c" ripple laminations and minor "b" laminations are common in graywacke layers in the lower part of the Seboomook (?) Formation.

The structure of the Carrabassett corridor is best interpreted as a warped and tightly appressed segment of earlier Acadian, presumably sub-isoclinal, or more open folds that originally trended more uniformly northeast. The structures display similar-type fold geometry where exposed in cross-section. In the southern part of the corridor the dominant tectonic grain defined by six subparallel fold axes is north to slightly east of north, except near Ira and Clay Brook Mountains where the trend swings strongly eastward. The plunge of folds is northerly and easterly, respectively, at shallow angles ranging up to 20°. From Little Poplar Mountain northward to the Bigelow range, fold axes have a strong east-west orientation with mild undulations of plunge. Limbs of folds are generally less tightly appressed, and

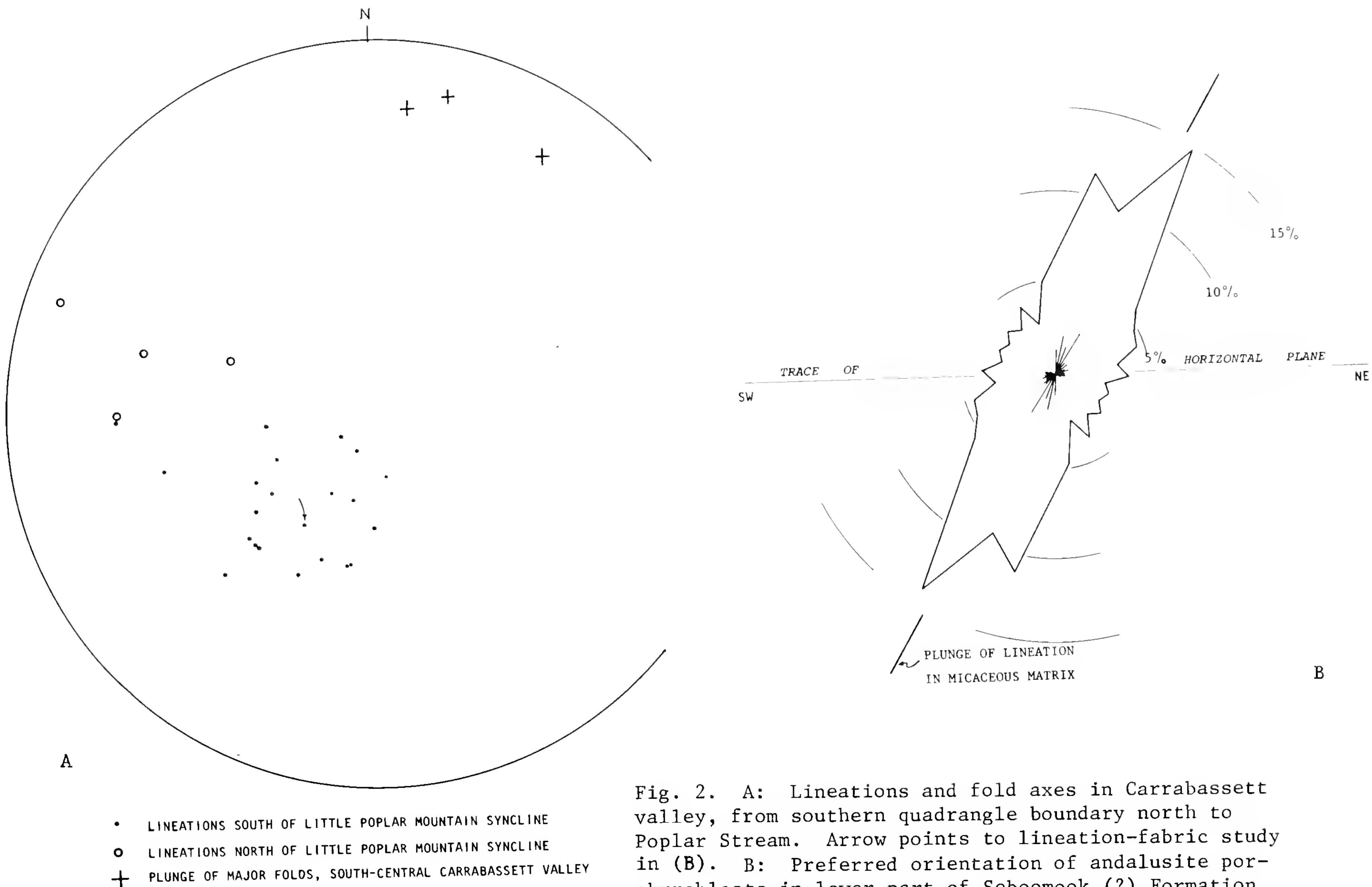


Fig. 2. A: Lineations and fold axes in Carrabassett valley, from southern quadrangle boundary north to Poplar Stream. Arrow points to lineation-fabric study in (B). B: Preferred orientation of andalusite porphyroblasts in lower part of Seboomook (?) Formation, Carrabassett valley (stop 6). Plane of diagram represents vertical foliation surface. Bar lengths at center indicate preferred orientations of boudinaged crystals.

fold amplitudes in the calc-silicate member of the Carrabasset Formation, and in the Seboomook (?) Formation are much decreased around the lower flanks of the Bigelow range.

The locus of structural change at Little Poplar Mountain also marks the center of a modified triangular syncline which is flanked by two gabbro bodies and by the westernmost lobe of the Lexington Quartz Monzonite.

Between Kingfield and the Little Poplar Mountain syncline, older rock units are exposed where the corridor is widest; correspondingly where the corridor is narrowest, younger units are exposed. It thus appears that the greatest vertical adjustment in response to late- and post-Acadian intrusion tectonics took place by flattening (crowding) between the intrusive boundaries where they form the most salient parts of the tectonic vice. Flattening was preceded or accompanied by subsidence, particularly during the intrusion of the Lexington batholith. Subsidence also was greatest where the lobes of bordering intrusions are more closely juxtaposed. Flattening and subsidence in the metasedimentary cover took place in response to the domical rise of the quartz monzonite while the nearby gabbroic bodies remained as semirigid buttresses. The orientation of least-principal stress is believed to have been steeply southwesterly-plunging at this stage of deformation in the south half of the corridor. Widely distributed occurrences of aligned, boudinaged andalusite pseudomorphs (see structural data, fig. 2) support this interpretation. Local modifications of stress-fields will be suggested at most of the stops.

The significance of lineation in the Carrabassett valley is intimately related to the metamorphic history, and is discussed below.

Metamorphism

The rocks of the corridor were metamorphosed in two thermal events of presumed late Acadian age. The first was associated with the emplacement of the Sugarloaf gabbroic massif and other mafic bodies of the area, was more intense, and affected a larger volume of rock. The second attended the emplacement and cooling of the Lexington batholith. Dikes and apophyses of granitic rocks cut the gabbros in numerous localities in the Little Bigelow Mountain quadrangle, and attest to relative ages.

Metamorphic relations in the corridor clearly indicate that the mafic intrusions caused much more intense, widespread metamorphism than did the granitic. The association of low pressure - intermediate facies progressions (fig. 4) with strongly developed tectonite fabrics bears much similarity to the Buchan area of the eastern Scottish highlands (Read, 1935) and to its facies series of metamorphism (Read, 1952; Craig, 1965). Similarities with the metamorphic aureoles of western Ireland are also to be noted below.

It has not been possible to determine to what extent early

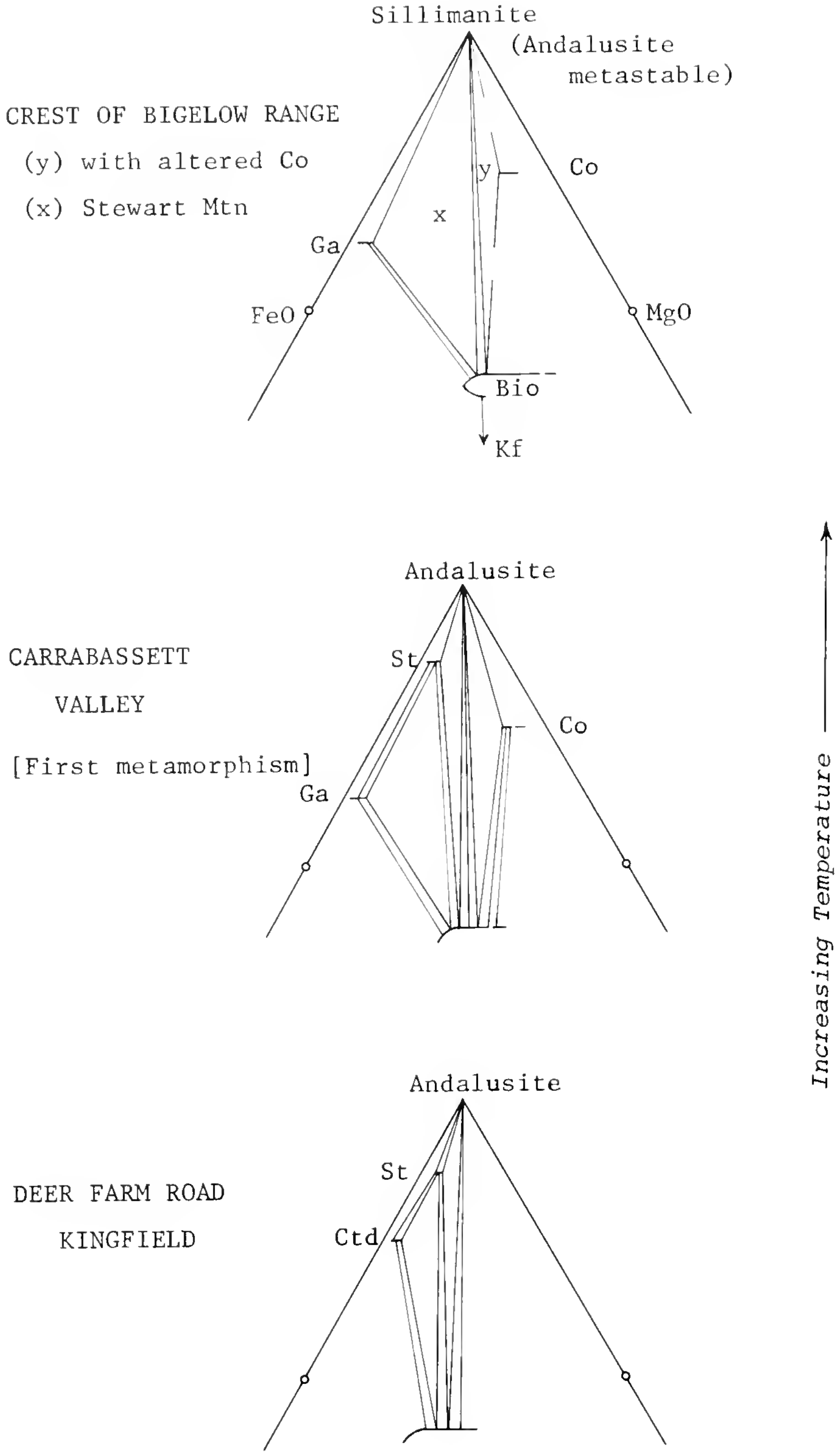


Fig. 3. Sequence of mineral assemblages in Carrabassett valley area from Kingfield township north to Bigelow Range. Diagrammatic; AFM projections from muscovite. Not shown is high-T, SiO_2-deficient assemblage cordierite-biotite-green spinel-orthopyroxene common in pelitic xenoliths in Sugarloaf massif.

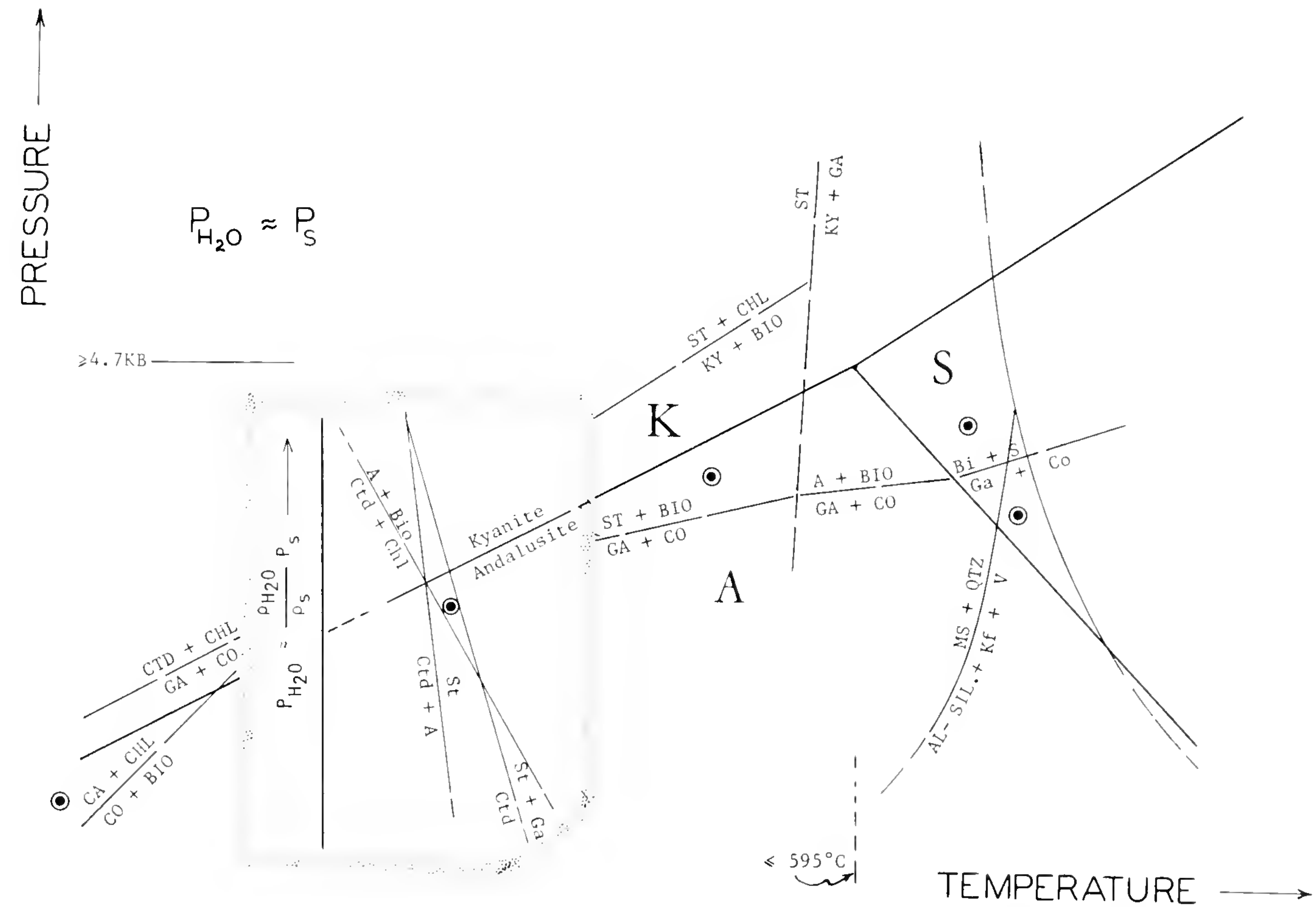


Fig. 4. Metamorphic zonal series for west half of Little Bigelow Mountain quadrangle and adjacent areas. Inset shows relative P_S-T location for chloritoid-staurolite-andalusite phyllite in west-central Kingfield township. Approximate locations and slopes of reaction boundaries from Albee (1965); coordinates of triple-point in Al_2SiO_5 system from Richardson and others (1969). Relative position of high-pressure limit for assemblage garnet + cordierite (+ biotite) modified from Hirschberg and Winkler (1968, fig. 1).

Acadian deformation had proceeded when the gabbroic magmas of the Sugarloaf massif and Flagstaff igneous complex intruded to the present level. Simple analogy with slate terrains to the northeast would imply that isoclinal folding had already been accomplished. However, shearing, protoclasis, and recrystallization of gabbroic rocks in marginal parts of the Sugarloaf massif and in mafic dikes and sills of the region suggest that Acadian folding continued up to the time of emplacement of the later granitic rocks of the region.

The emplacement of the Sugarloaf gabbroic massif initiated widespread metamorphism of the Siluro-Devonian rocks of the Carrabassett valley to schists and gneissic granofelses collectively displaying a Buchan type of facies series. At intermediate grades of this metamorphism, two sets of assemblages occur in metapelites. The pelitic rocks in west central Kingfield township contain the assemblages andalusite-staurolite-biotite, and staurolite-chloritoid-biotite in a zone of strong penetrative deformation where H_2O pressure was probably much less than mean lithostatic pressure. Elsewhere in the Carrabassett valley the pelites contain the closely equivalent, probably slightly higher temperature assemblages andalusite-cordierite-biotite, andalusite-staurolite-biotite, and staurolite-garnet-biotite (compare, fig. 3). All pelites studied in the Carrabassett valley contain excess muscovite and quartz, but the second group of assemblages denote parageneses in which H_2O pressure may have more nearly equalled lithostatic pressure.

Relative sizes of andalusite, staurolite, and garnet, and their mutual host-inclusion relationships are thought to represent bulk compositional control on crystallization rates, rather than indicating the timing of the beginning and cessation of their crystallization.

At highest metamorphic grade, biotite remained stable, and is accompanied by almanditic garnet and sillimanite.

(Low ridges and hills near the contact in the vicinity of Grindstone and Shiloh ponds mark the widest part of the sillimanite zone on the eastern side of the Sugarloaf massif. Rocks of this zone do not crop out near the highway or river. East of the river, sheared relict andalusite-bearing schist is found within a few yards of the gabbro, and suggests that in the central part of the valley the contact is faulted, with the gabbro thrust over the inner part of the aureole.)

Cordierite has not been found in the SiO_2-saturated pelitic rocks of the sillimanite zone (southwestern part of the map area). Corresponding behavior has also been noted by Gribble (1966) in the aureole of the Haddo House norite, Scotland. Rafts of roof-pendants of pelitic rock well within the Sugarloaf massif, however, (such as capping the summit of Owl's Head) consist of abundant cordierite accompanied by green (hercynite-rich?) spinel, hypersthene, and plagioclase. This suggests that the ubiquitous silica-oversaturated pelites were desilicated at the upper contact, and were engulfed by the gabbroic intrusion. Leake and Skirrow (1960) also proposed desilication for quartzose pelites in contact with mafic and ultramafic rocks in western Ireland.

The andalusite produced during this first stage of metamorphism in the Carrabassett valley is chiastolitic, is of large porphyroblastic dimensions, and exhibits a weak, preferred orientation within the plane of schistosity.

The stage during which the andalusite became preferentially oriented to produce the weak linear fabric is not bracketed with certainty. It could have been initiated during the first stage of prograde dynamothermal metamorphism accompanying the emplacement of the gabbroic massif, but it clearly was associated with the second stage of metamorphism and deformation during the emplacement of the Lexington batholith. Pitcher and Read (1963) proposed that a weak linear fabric in the Dalradian schists of the late Caledonian Thorr aureole of western Ireland, marked in part by aligned andalusites, developed by mimetic growth of andalusite parallel to "well-marked puckers produced by the new strain-slip cleavage;" (ibid., p. 268; 273). In the Carrabassett valley, on the contrary, Gleitbrett structure and shear schistosity (cf. Whitten, 1966, p. 137-140) appear to have been produced late in the metamorphic-tectonic history when andalusite, staurolite, and garnet were abraded along strain-slip surfaces, andalusite crystals were boudinaged, and all three aluminosilicates were largely replaced by muscovite and chlorite. Crinkling and grooving are observed on the macroscopic scale to originate at, and tail away from porphyroblast idiomorphs, fragments, and other resistant mineral grains. The smallest-scale crinkles diverge with the foliation around porphyroblasts. Therefore it seems that the crinkling and grooving of the schistosity surfaces owes its origin to drag produced by differential movement between micaceous matrix and resistant, chiefly porphyroblastic mineral fragments.

The metamorphism associated with the batholithic intrusion produced prograde andalusite-, cordierite-almandite-, and sillimanite-almandite-bearing gneissic granofelses near the batholith contacts, but did not produce staurolite. The widths of these zones around the batholith are narrower than those associated with the gabbroic massif (even around the east contact where no polymetamorphic complications exist). Crystal size of andalusite is smaller, and where the second thermal gradient overlapped the metamorphic zones of the first, retrograde reactions commonly accompanied partial phyllonitization. The less widespread, more locally confined prograde metamorphism that accompanied the emplacement of the Lexington batholith here can be explained in part, in terms of the batholith's emplacement into rocks of the corridor which had already undergone prograde dehydration reactions during the intrusion of the Sugarloaf massif. If, as seems certain, the fluids from such earlier reactions were driven out of the volume of rock now underlying the corridor, the source of fluids for the formation of retrograde chlorite and muscovite in these rocks must be sought from the newly-formed aluminosilicate-, garnet, and cordierite-bearing gneissic granofelses adjacent to the batholith, and perhaps from water expelled from crystallization of hydrous granitic magma. Evidence for and against this latter possibility will be reviewed during the trip.

Chiastolite porphyroblasts produced in stage 1 were rotated, boudinaged, abraded, and altered to pseudomorphs consisting mainly of fine-grained white mica. Staurolite, cordierite, and garnet were largely altered to chlorite-white mica pseudomorphs.[2] The planar fabric diagram (fig. 2) indicates the degree of parallelism of andalusite pseudomorph alignment and boudinage with the locally pervasive "crinkle" lineation in the schistose matrix. The degree of rotation and destruction of porphyroblasts corresponds to the intensity of the southwesterly-plunging lineation.

The southwesterly-plunging lineation is widespread in the pelitic schists of the Carrabassett corridor, and is perhaps its most undeviating structural feature. It does not bear a fixed relationship to plunge of major fold axes, as we will observe, yet it approximates a tectonic a lineation at many localities in the south half of the corridor where major folds plunge to the north or northeast. Its origin is ascribed to late tectonic, inhomogeneous, nonaffine shearing, and is believed to represent the final structural development in the schists of the corridor, attending late-stage doming of the Lexington batholith. The shear-dislocation of crests and troughs of minor folds may, but not necessarily, have taken place concurrently with the development of the lineation.

Large bulbous masses of quartz, and ptygmatically folded quartz veins are common in the southern part of the corridor. The axes of the ptygmatic structures generally plunge southwesterly. Where armored by quartz, pink, fresh andalusite and occasionally blue cordierite have been found. Swarms of chiastolite, preserved as white mica pseudomorphs, and abundant tourmaline commonly line the walls of the quartz bulbs and veins. This rather varicose geometry of the quartz-congested plumbing system may represent deformation of previously more planar structures, but more likely, simultaneous growth of the structures and sweating out of these constituents from the schists in a mileau of ductile deformation, which began in the first dynamothermal event and ceased during the second. There is no evidence that this hydrothermal system or the source of boron can be genetically traced to the crystallization of mafic or sialic magmas; rather, the decreasing concentration of boron (tourmaline: axinite (?)) into the mafic and granitic plutons suggests that it originally accumulated in the pelitic sediments at the time of deposition.

---

2 A petrographic study of the pseudomorph-bearing schistose layers at stop 6, carried out by P.T. Lyttle, shows that the Al_2O_3- constant reaction:

$$\text{Biotite}_a + \text{Andalusite} + H_2O \rightarrow \text{Muscovite} + \text{Biotite}_b + \text{Chlorite} + \text{Quartz}$$

may require at least 0.1 mole of K^+ introduced per mole of andalusite altered to muscovite.

Microprobe analysis across a large chiastolite pseudomorph containing a core of fresh andalusite shows that a thin, discontinuous zone of pyrophyllite separates part of the fresh core from surrounding muscovite.

References

Albee, A.L., 1965, A petrogenetic grid for the Fe-Mg silicates of pelitic schists: Am. Jour. Sci., v. 263, p. 512-536.

Boucot, A.J., 1969, Geology of the Moose River and Roach River synclinoria, northwestern Maine: Maine Geol. Survey Bull. 21, 117 p.

Craig, G.Y., 1965, The geology of Scotland: Oliver & Boyd, Edinburgh, 556 p.

Gribble, C.D., 1966, The thermal aureole of the Haddo House norite in Aberdeenshire: Scottish Jour. Geology, v. 2, p. 306-313.

Hirschberg, Achim, and Winkler, H.G.F., 1968, Stabilitatsbeziehungen zwischen Chlorit, Cordierit und Almandin bei der Metamorphose: Contr. Mineralogy and Petrology, v. 18, p. 17-42.

Leake, B.E., and Skirrow, Geoffrey, 1960, The pelitic hornfelses of the Cashel-Lough Wheelawn intrusion, County Galway, Eire: Jour. Geology, v. 68, p. 23-40.

Pitcher, W.S., and Read, H.H., 1963, Contact metamorphism in relation to manner of emplacement of the granites of Donegal, Ireland: Jour. Geology, v. 71, p. 261-296.

Read, H.H., 1935, The gabbros and associated zenolithic complexes of the Haddo House district, Aberdeenshire: Quart. Jour. Geol. Soc. London, v. 91, p. 591-635.

__________, 1952, Metamorphism and migmitisation in the Ythan Valley, Aberdeenshire: Edinburgh Geol. Soc. Trans., v. 15, p. 265-279.

Richardson, S.W., M.C. Gilbert, and P.M. Bell, 1969, Experimental determination of kyanite-andalusite and andalusite-sillimanite equilibria; the aluminum silicate triple point: Am. Jour. Sci., v. 267, p. 259-272.

Whitten, E.H.T., 1966, Structural geology of folded rocks; Rand McNally, Chicago, 663 p.

Trip A-2

Road Log

Trip will start from Rangeley at 8:00 a.m. Assemble near Rangeley Inn. We will proceed directly to Kingfield; all localities for the trip are in the northwestern part of the Kingfield quadrangle and southwestern part of the adjoining Little Bigelow Mountain quadrangle. J. Raabe will lead the discussion on stops 1 - 2a; G.M. Boone will lead the remainder of the trip.

Mileage

0.0	Proceed E on Rt. 4 to Madrid and Phillips; at Phillips proceed NE on Rt. 142 to Kingfield.
38.0	Turn right, at termination of Rt. 142 in Kingfield, approximately 0.2 mile SE on Rt. 16 to bridge over the Carrabassett river. Park beyond E end of bridge.
38.2	STOP 1. Madrid formation. Medium- to thickly-bedded calc-silicate layers, the exposure here in the river being somewhat more calcic than is regionally typical of the Madrid. Direction in which the nearly vertically dipping beds face is largely indeterminate here. The assemblage plagioclase-actinolite-epidote±biotite±garnet is typical of the low to intermediate regional metamorphic grade of this part of the Kingfield quadrangle. Major aspects of relict parallel bedding are preserved, although most of the small-scale textures are metamorphic. Lineation plunges steeply.
38.4	Return across bridge and proceed north on Rt. 16-27. We cross the unexposed north contact of the Madrid formation approximately a mile north of the village limits.
39.7	Turn left onto Deer Farm road (sign at corner); proceed 1.0 mile from turn-off to road-side outcrops of andalusite-staurolite phyllite of presumed lower Devonian age.
40.7	STOP 2. Cleavage and bedding here dip essentially vertically. This locality is more than a mile E of the ductile shear zone, but its essential features and relations to the structures observed here will be discussed (see preceding text by Raabe). The pelite here reflects an intermediate grade of metamorphism, with layers 3-4 mm thick containing the assemblages biotite-staurolite-andalusite, and biotite-staurolite-chloritoid (see discussion by Boone).

41.7 STOP 2a, tentative. Weather permitting, we will proceed to the top of the hill where we will briefly point out some of the larger-scale features of the southern part of the Sugarloaf gabbro and its roof-rocks, before retracing our route back to the Carrabassett valley and Rt. 16-27. Poorly exposed glacial pavement in the field immediately east of the road at the crest of the hill exposes lower Devonian (?) fine-grained muscovite-chlorite phyllite.

Return to Rt. 16-27 and proceed north approximately 2.15 miles.

45.9 STOP 3. We have crossed the quadrangle boundary 0.5 mile into the Little Bigelow Mountain quadrangle, and the outcrop location is roadside just north of the circled route numbers on the topographic map. The dominantly pelitic sequence extending northward through the valley has been divided into two formations (see text) which are collectively referred to as the Seboomook (?) formation along strike southwestward. We are, then, here in the lower member (approximately 3500 ft. thick) of the Carrabassett formation. The member is dominantly "massive" throughout most of its extent, but we see here one of several well-bedded parts, made so by locally abundant interbedded quartz wacke.

NO HAMMERING HERE.

These well-bedded portions superficially resemble the Perry Mountain formation of upper Silurian (?) age, seen on trip A-1. Comparison shows, however, that the Perry Mountain is consistently well-bedded, and is more potassic (muscovite is much more abundant) in its pelitic layers. Ratio of Fe/Mg seems generally higher, but total Fe, Mg, is lower in the Perry Mountain than in the pelites of the Carrabassett or Seboomook formations.

The structures displayed here have lured the attentions of many. Locally known by a legion of geologists as "the butterfly fold", it is the femme fatale of Carrabassett tectonics. Interpretations of its structural history have also been legion, but I offer the following provocation: In the later stages of vertical stretching that accompanied shearing and flattening of folds, we witness here a zone of greater-than-normal vertical elongation, with a local quasi-uniaxial stress field, the orientation of least principal compressive stress plunging steeply southwestward, parallel to the lineation. Secondary, steeply plunging folds were created by a puckering of the section inward, the secondary axial surfaces somewhat radially disposed toward the axial zone of maximum elongation. We see this local geometry planed off near a zone of necking or boudinage, in which the central quartz bulbs represent local pressure shadows.

Although quartz-filled tension-gashes abound in the quartz wacke beds, some may wish to argue for initial development of this structure in post-depositional slumping in the sedimentary environment. "Tectonic inclusions" of quartz wacke surrounded by pelite may represent semi-coherent sand lumps rafted into place during a period of redeposition in the sedimentary trough.

Retrograde recrystallization from andalusite-cordierite grade to form biotite-muscovite-chlorite, will be discussed. Note the abundant tourmaline associated with the quartz bulbs.

Proceed north 1.7 miles.

47.6 STOP 4. This laterally extensive, vertical rock face exposes a biotitic metaquartzite member, representing one of several thick, discontinuous clastic depositional wedges in the central part of the Carrabassett formation. It is preserved here in a recumbent, doubly-plunging syncline. Here the style of deformation in a much more resistant lithic type can be observed. Medium to thick, well-graded beds face normally and dip moderately westward at the south end of the exposure. The axial surface descends close to or beneath the road-level at the north end of the outcrop; here minor recumbent folds can be seen in the axial zone. Garnetiferous aplite dikes and quartz-pyrite veins cut the structure.

Proceed north approximately 0.2 mile.

47.8 STOP 5. Location is 0.5 mile N of Kingfield-Jerusalem township line. Outcrops here display the typically "massive" characteristics of relict bedding, (or total lack of bedding) in the lowest member of the lower Devonian (?) section. The andalusite schist has been sheared and retrograded. There are well-displayed examples here of ptygmatic quartz veins bordered by clusters of large chiastolite pseudomorphs.

Proceed north approximately 1.2 miles, to sharp bend in road paralleling the abrupt bend in the Carrabassett River.

49.0 STOP 6a. We have crossed over the uppermost (calc-silicate) member of the Carrabassett formation, which crops out only at higher elevations on each side of the river. Stops 6a and b are located in the lowermost part of the Seboomook (?) formation, here used in a more restrictive sense than in the region to the southwest. An additional estimated thickness of 1800 feet of the Seboomook (?) is preserved in this western half of the quadrangle.

This and other sections of the formation are consistently graded-bedded, with graded sets ranging from 2 inches to a foot thick. Basal graywacke parts of the sets show wavy bedding, cross-bedding, and rarely, foreset terminations at their bases, which produce a false-crossbedded appearance with respect to the top of the subjacent metapelite.

Most of the section here faces SE, but we are in the axial zone of a major syncline and "top" reversals across subsidiary folds throughout this zone are common, as can be seen here to advantage. Also well exposed here and at stop 6b is the style of widely spaced shearing and rotation between and within minor fold segments.

A "porphyroblast-drag" lineation, to be discussed, plunges steeply in the usual SW direction; its relation to fold axes is best observed and discussed here as well as at the next stop (see comments in text of accompanying article). Small scale, quartz-rich streaky segregations occur parallel to bedding in the metapelite in the northern part of the outcrop; its control, however, is cleavage and not bedding, as observed in many locations where bedding crosses cleavage at high angles.

Well preserved idiomorphs of largely altered staurolite and chiastolite represent the first recorded phase of metamorphism associated with emplacement of the Sugarloaf massif.

Proceed around bend in road, 0.3 mile.

49.3 STOP 6b. The exposure of folds in the three dimensions: cleavage, the near-horizontal, and the normal to fold-axis plunge, is best afforded here, even though structural setting and stratigraphic position are similar to 6a. The moderate preferred orientation of andalusite is also well displayed, and the fabric diagram is based on measurements from this locality.

The itinerary from stop 3 to the present location has been across a uniform zone of prograde metamorphism related to the gabbro, in which biotite-andalusite-staurolite (±garnet ±cordierite) characterized the assemblages. For reasons reviewed in the accompanying article, higher-grade assemblages are not evident between here and the gabbro contact crossed by the highway 0.8 mile northward.

The second phase of metamorphism, keyed to the emplacement of the Lexington batholith, is also uniformly represented over this same section of the itinerary by retrograde metamorphism yielding brown biotite, abundant chlorite, and muscovite. Varied degrees of shearing and phyllonitization are visible in outcrops, but particularly in thin sections of these rocks.

One may note here, as in previous stops, that muscovite and feldspar are notably absent from the interior parts of quartz veins. Although there is some suggestion, from modal analysis and calculations involving standard metamorphic biotite $K/Al^{VI} + Al^{IV}$ ratios, that potassium metasomatism is necessary in the metapelites at outcrop localities 6a and b to account for the muscovite replacement of andalusite and staurolite, the calculated amount is slight (see accompanying article) and cannot be established with certainty. All the K needed for muscovitization of andalusite and staurolite may have been supplied by chloritized biotite.

Proceed north 1.6 miles.

50.9 STOP 7. This is the last scheduled stop of Trip A-2, and the location on the topographic map is Spring Farm, B.M. 789. LUNCH. (Weather permitting, we can proceed ahead to the small field at the intersection of the old road on right, just north of the bridge.)

We see here the outer part of the Sugarloaf gabbroic massif. Characteristic of much of the massif, the gabbroic rocks here are altered and in places sheared. Massif-wide observations and study of exposed contacts suggest that the massif was emplaced and became largely solidified before regional folding and tectonic activity ceased.

Preserved here in part, nonetheless, are olivine-bearing, pyroxene-rich layers and plagioclase-rich layers in a rhythmically-layered sequence. (Not all of the massif is so layered.) The massif can be said to have tholeiitic affinity if the presence of both ortho- and clinopyroxenes is taken as a standard criterion. The igneous rocks of the massif have not been systematically mapped or studied.

Note that the strike of the layering is perpendicular to that of the contact, and the layering here characteristically dips gently northward.

Return 8.5 miles to Kingfield (Farmington is 22 miles due south on Rt. 27); return to Rangeley by turning right onto Rt. 142 to Phillips; Rt. 4 Phillips to Rangeley.

Geology of the Lower Paleozoic Rocks in the Boundary Mountain Anticlinorium[1]

By David S. Harwood,[2] John C. Green,[3] and Charles V. Guidotti[4]

Introduction

The Boundary Mountain anticlinorium (Albee, 1961) in northern New Hampshire, west-central Maine, and adjacent Québec consists of a core of pre-Silurian rocks overlain to the northwest and southeast, respectively, by Silurian and Devonian rocks in the Connecticut Valley-Gaspé synclinorium (Cady, 1960) and the Merrimack synclinorium (Osberg, Moench, and Warner, 1968). This trip will examine the highlights of the stratigraphy, structure, and metamorphism of the rocks in the anticlinorium, beginning in the Silurian rocks on its southeast flank near Rangeley, Maine, and proceeding westward across its core to a synclinal inlier of Silurian rocks near Parmachenee Lake (fig. 1). Emphasis will be given to the basal Silurian clastic rocks in two widely separated parts of the region and to some of their possible source rocks in the core of the anticlinorium.

Before the mid-1950's, the geology of this remote woodland area, shown in figure 1, was known only through the reconnaissance studies of Hitchcook (1877) and Logan (1863). During the past 15 years, however, detailed mapping in the Errol, Second Lake, and Moose Bog quadrangles by Green (1964, 1968), in the Dixville quadrangle by Hatch (1963), in the Oquossoc quadrangle by Guidotti (Maine Geol. Survey, open-file report), and in the Cupsuptic and Arnold Pond quadrangles by Harwood (1966, 1968) has extended the classical New Hampshire stratigraphic sequence worked out by Billings (1937, 1956), into northernmost New Hampshire and west-central Maine. McGerrigle (1935) and Marleau (1957, 1959, 1968) have mapped the rocks in the adjacent part of Québec.

Stratigraphy

The stratigraphy of the field-trip area is briefly summarized below; more detailed descriptions of the rocks are available in the references cited:

<u>Aziscohos Formation</u>: (Green, 1964). Lower part is predominantly carbonaceous rusty-weathering schist and phyllite composed of quartz, plagioclase, muscovite, biotite, and chlorite with garnet, staurolite, andalusite, and sillimanite present in certain rocks at the appropriate grades of metamorphism. Upper part is green, silvery-green to gray phyllite and schist characterized by abundant stringers and pods of milky-white quartz. In addition to these principal rock types, the formation contains about 10 percent of biotite-quartz-plagioclase gneiss, carbonaceous quartzite, calc-silicate rock, and quartz-spessartite rock.

---

[1]Publication authorized by the Director, U. S. Geological Survey.

[2]U. S. Geological Survey, Washington, D.C.

[3]University of Minnesota, Duluth.

[4]University of Wisconsin.

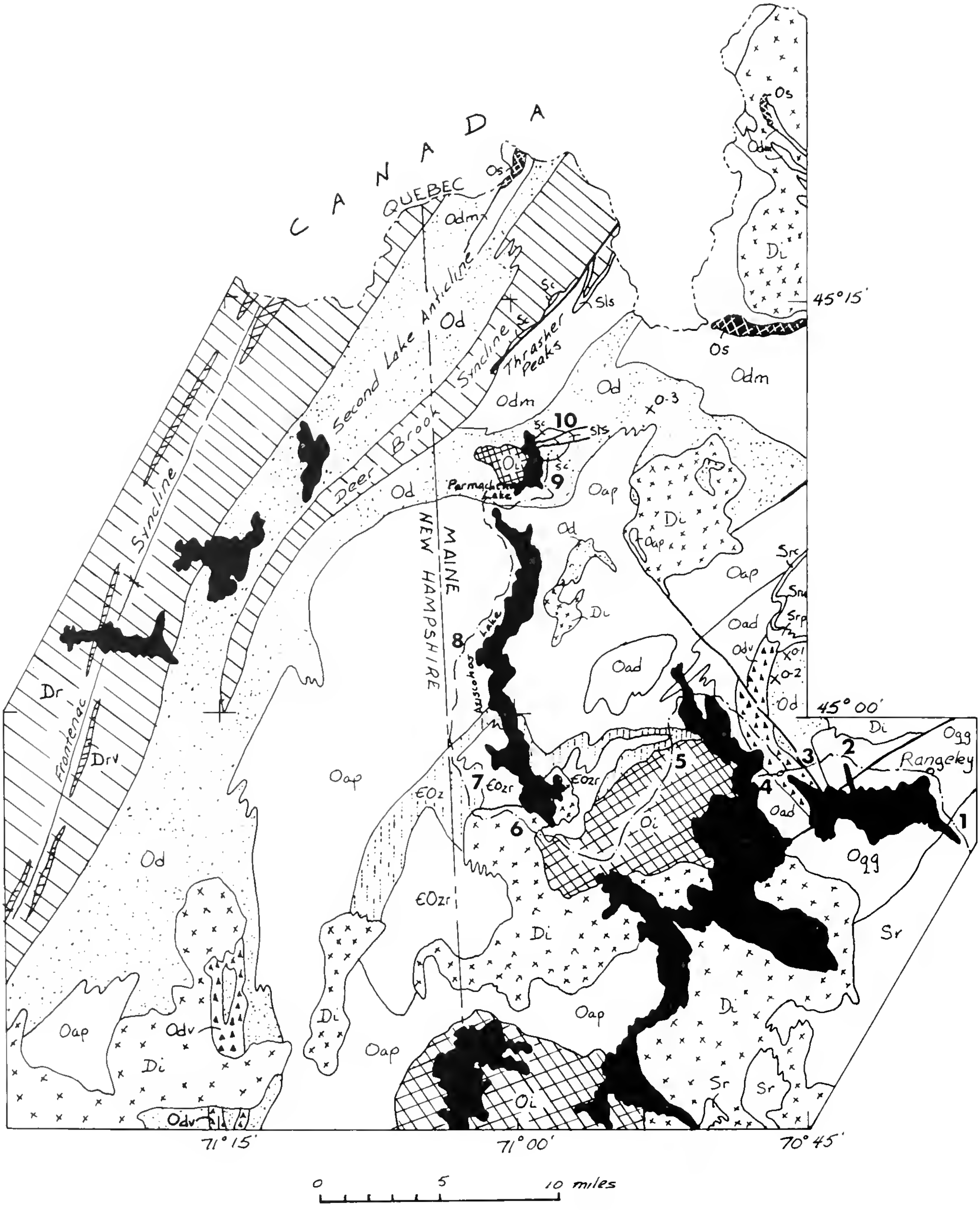


Figure 1 - Generalized geologic map of the southwestern part of the Boundary Mountain anticlinorium

EXPLANATION

Lower Devonian

[illegible] rocks

Dr- [illegible] and quartzite of Gile Mountain, [illegible], Frontenac, and Compton Formations; contains mafic meta-volcanic rocks, Drv

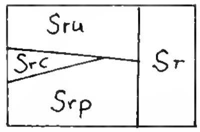
Upper Silurian

Unnamed Silurian rocks

Sls-calcareous slate and limestone

Sc -quartz-pebble and polymict conglomerate

Lower Silurian



Rangeley Formation

Sr-parts A,B, and C and small amount of Perry Mountain Formation south of 45th parallel (see Trip A-1)

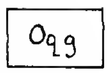
Sru-gray slate, quartzite, and minor polymict conglomerate lenses

Src-quartz-pebble conglomerate and quartzite

Srp-polymict conglomerate and quartzite

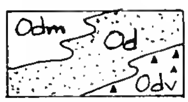
UNCONFORMITY

Upper Ordovician(?)



Quimby and Greenvale Cove formations; undivided

Middle Ordovician

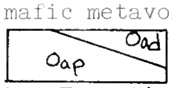


Dixville Formation

Odm-Magalloway member; feldspathic graywacke and subordinate slate, granule conglomerate, and metavolcanic rocks

Od -black slate, feldspathic quartzite, and layers of predominantly mafic metavolcanic rock

Odv-predominantly mafic metavolcanic rock

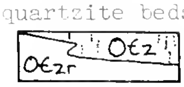


Albee Formation

Oad-green, gray, black, and purplish-gray slate with minor arenaceous beds

Oap-green, gray, and red slate or silvery-gray schist with "pinstripe" quartzite beds

Upper Cambrian and Lower Ordovician



Aziscohos Formation of Green (1964)

O€z-green to silvery-gray phyllite and schist with quartz pods and stringers

O€zr-black to dark gray carbonaceous phyllite and schist, minor arenaceous and metavolcanic rocks

Devonian intrusive rocks

Ordovician intrusive rocks

Orodovician serpentinite and related mafic rocks

O-1
X
Fossil locality

Fault

.4

Field trip route showing stops

Albee Formation: (Billings, 1937, 1956: see also Green 1964, 1968; Hatch, 1963; Harwood, 1966). South of the 45th parallel the Albee is primarily silvery-gray to greenish- or brownish-gray schist that contains distinctly laminated "pinstripe" quartzite and feldspathic quartzite beds. North of the 45th parallel it is predominantly green to gray-green slate and phyllite with the same diagnostic arenaceous rocks interbedded in variable proportions. In the Cupsuptic quadrangle and adjacent parts of the Oquossoc quadrangle, the Albee contains red slate with interbedded "pinstripe" quartzite and green, gray and purplish-gray slate with abundant quartz stringers and scarce arenaceous beds.

Dixville Formation: (Green, 1964, 1968; Hatch, 1963, Harwood, 1966; Harwood and Berry, 1967). The Dixville consists predominantly of black to dark-gray carbonaceous rusty-weathering schist, phyllite, and slate with interbeds of feldspathic metasandstone and locally calcareous lithic graywacke and minor conglomerate. Mappable patches and layers of mafic and felsic metavolcanic rocks are present at different stratigraphic levels within the black slate. In the northern part of the Second Lake and Cupsuptic quadrangles the upper part of the formation is a heterogeneous collection of feldspathic graywacke, arkosic conglomerate lenses, black, green, and purplish-gray slate, and lenticular patches of metavolcanic rocks mapped as the Magalloway Member (Green, 1968; Harwood 1966). The Magalloway Member is continuous across the international boundary with the Arnold River Formation of Marleau (1968).

Quimby Formation: (Moench, 1969). This formation consists of a lower metagraywacke member that contains interbedded conglomeratic metagraywacke, gray to black slate, and felsic metavolcanic rocks and an upper metashale member that contains cyclically bedded metagraywacke and metashale interlayered in equal proportions.

Greenvale Cove Formation: (Moench, 1969). Light- and dark-gray metamorphosed shale, siltstone, and sandstone are interlayered on a fine scale to produce a characteristically banded rock. The formation locally contains calc-silicate rock and biotite granofels (see Moench and Boudette, Trip A-1, this volume).

Rangeley Formation: (Osberg, Moench, and Warner, 1968; Moench and Boudette, this volume). According to R. H. Moench (written commun., 1970) the polymict conglomerate at stop 1 contains well-rounded clasts of felsic and mafic metavolcanic rocks, metamorphosed lamprophyric rocks, vein quartz, metashale, metasiltstone, quartzite, crinoid-bearing limestone, and various granitic rocks, including distinctive medium-grained granodiorite and quartz diorite with blue quartz.

In the Cupsuptic quadrangle the polymict conglomerate contains abundant platy fragments of black slate, minor amounts of green slate, boulders of coral-bearing limestone, as well as the clasts listed above for stop 1. Fossil fragments in the matrix and in the limestone boulders in the Cupsuptic area suggest a Silurian age (Harwood and Berry, 1967). Quartz-pebble conglomerate that overlies the polymict conglomerate in the Cupsuptic quadrangle is considered equivalent to

similar rocks that contain Early Silurian (C_3 -C_5 of the Late Llandovery) fossils in the Kennebago Lake quadrangle, reported by E. L. Boudette (U. S. Geol. Survey, 1965, p. A-74). (See Moench and Boudette, this volume).

Unnamed Silurian rocks at Parmachenee Lake: (Harwood, 1966, Harwood and Berry, 1967). Gray argillite and tuff interfingers with and underlies quartz-pebble conglomerate containing interbedded purple slate, gray feldspathic quartzite, and one lens of polymict conglomerate. The conglomerate is succeeded by tan-weathering, mottled gray and white well-bedded argillaceous limestone which, in turn, is overlain by tan to greenish-tan pit-weathering fossiliferous arenaceous limestone that contains Late Silurian (Pridoli age) fossils, according to A. J. Boucot (written commun., 1968).

Unnamed Silurian rocks near Thrasher Peaks: (Green, 1968; Harwood, 1966). Polymict conglomerate composed of pebbles and cobbles of feldspathic graywacke (similar to the Magalloway Member), gray metasiltstone and metasandstone, and mafic metavolcanic rocks occurs as lenses on Thrasher Peaks (fig. 1). Northeast of Thrasher Peaks, gray calcareous slate and lenses of limestone and limestone conglomerate of probable Ludlow age (Green, 1968, p. 1615) rests unconformably on the Magalloway Member of the Dixville.

Undivided Silurian and Devonian rocks on the northwest flank of the anticlinorium: (Green, 1964, 1968; Hatch, 1963; Marleau, 1968; Harwood, 1969a). The youngest metasedimentary rocks on the northwest flank of the Boundary Mountain anticlinorium are gray slate and phyllite containing variable amounts of interbedded gray quartzite and lenses of mafic metavolcanic rocks. These rocks have been assigned to the Gile Mountain Formation (Hatch, 1963), the Frontenac Formation (Marleau, 1968: Green, 1968; Green and Guidotti, 1968), and the Seboomook and Compton Formations (Green, 1968; Marleau, 1968) all of known or inferred Early Devonian age (see Boucot, 1961). The Devonian rocks in the Second Lake, Cupsuptic, and Arnold Pond quadrangles are separated locally from the Silurian or pre-Silurian rocks by lenticular units of white to greenish-gray felsic metavolcanic rocks (Harwood, 1969a; Green, 1968, Prospect Mountain tuff-breccia) and patches of red slate or green slate and calcareous metasiltstone of Late Silurian or Early Devonian age, according to Harwood (1969a).

Correlation and age of the pre-Silurian rocks

Graptolites at locality O-1 (fig. 1) establish a late Middle Ordovician age (Harwood and Berry, 1967) for the black slate assigned to the Dixville Formation in the southeastern part of the Cupsuptic quadrangle. The descending lithologic sequence of black slate-mafic metavolcanic rocks-green slate and "pinstripe" quartzite is the same as the classical New Hampshire sequence of Partridge-Ammonoosuc-Albee (Billings, 1937) and, furthermore, the Albee can be traced continuously from the Cupsuptic quadrangle to its type area near Littleton, New Hampshire.

Thus, a late Middle Ordovician age seems reasonably certain for the Partridge Formation; the Ammonoosuc Volcanics and the Albee must be Middle Ordovician or older.

The belt of Dixville north and west of the Albee has not been unequivocally dated as yet, but it must be pre-Silurian because it unconformably underlies Silurian rocks near Thrasher Peaks and Parmachenee Lake and contains Protospongia sp. of Cambrian or Ordovician age at locality O-3 (fig. 1). Green (1964, 1968) and Hatch (1963) correlated the Dixville with the Partridge Formation and Ammonoosuc Volcanics on the basis of lithologic similarity and stratigraphic position above the Albee. In the belt of rocks, however, the mafic metavolcanic rocks are distributed throughout the black slate, the latter commonly being in contact with the Albee Formation. Harwood and Berry (1967) correlated the Dixville with the graptolite-bearing slate at locality O-1 on the basis of lithologic similarity, stratigraphic position, and the presence of Middle Ordovician fossils in similar rocks on strike to the northeast.

Billings (1956, p. 98) concluded that the Albee was Middle Ordovician or older through correlation with the Moretown Member of the Missisquoi Formation of Doll et al. (1961) in eastern Vermont. Recently Zen (1967, p. 55) has correlated the Moretown Member and part of the underlying Stowe Formation with the B and C members of the Poultney Slate (Theokritoff, 1964) in the northern Taconic Mountains. This part of the Poultney contains late Early Ordovician to early Middle Ordovician fossils (Berry, 1961) and bears striking lithologic similarity not only to the Stowe Formation and the Moretown Member but to the non-carbonaceous upper part of the Aziscohos Formation and the Albee. Green (1964), in fact, correlated the upper Aziscohos with part of the Stowe Formation and the Albee with the Moretown Member. Thus, an Early to Middle Ordovician age is tentatively assigned the upper Aziscohos and Albee. Green (1964) suggested a possible correlation between the lower carbonaceous part of the Aziscohos and the lower part of the Stowe Formation and perhaps the underlying Ottauquechee Formation of eastern Vermont. If this is correct, a Middle Cambrian to Early Ordovician age is implied for the lower part of the Aziscohos Formation.

Tectonic Events

Although other stratigraphic and structural interpretations are possible for parts of the region, the major tectonic events are fairly well documented and are presented chronologically in the following outline:

1. Pre-Silurian deformation: Gently dipping Silurian rocks truncate the steeply dipping Albee-Dixville contact in the southeastern part of the Cupsuptic quadrangle and rest upon the Dixville near Parmachenee Lake. Farther to the northwest, Silurian and Devonian rocks truncate the contact between the

black slate and Magalloway members of the Dixville. These relationships indicate that the pre-Silurian rocks were folded, uplifted, and eroded during the Taconic orogeny, which can be dated in this area as later than zone 12 of the late Middle Ordovician and before C_3-C_5 time of the late Llandovery (Early Silurian). Lineations in the pre-Silurian rocks are markedly steeper and differ in direction from those in the Silurian and Devonian rocks in the Second Lake quadrangle (Green, 1968), also implying considerable Taconic folding. No pervasive cleavage can be unequivocally related to this pre-Silurian folding. The map pattern and the orientation of minor folds, however, suggests that early folding was about west- to northwest-trending, vertical or steeply dipping after the Taconic folding. Intrusion of granodiorite and granite plutons and smaller bodies of ultramafic rock preceded and possibly accompanied the early deformation.

2. Post-Early Devonian deformation: The major period of deformation, affecting all the stratified rocks in the area, is related to the Middle Devonian Acadian orogeny. The Silurian and Devonian rocks were deformed into relatively open to isoclinal northeast-trending folds that plunge at shallow angles generally to the northeast but locally to the southwest. In contrast to this, the pre-Silurian rocks contain steeply plunging to vertical northeast-trending isoclinal folds superimposed on the steeply dipping limbs of the Taconic folds. Complex refolded folds and folded lineations are common in the pre-Silurian rocks and are particularly well shown in parts of the well-bedded Albee Formation. Foliated to massive irregularly shaped masses of biotite-muscovite quartz monzonite intruded the rocks south of the 45th parallel after the major folding. North of the 45th parallel, masses of nonfoliated biotite-muscovite and biotite-hornblende quartz monzonite apparently worked their way upward by stoping and block caving after the main period of folding. Adjacent to these later intrusive rocks and in scattered areas removed from visible late intrusive rocks the country rocks are deformed by vertical, north-trending slip cleavage.

3. Post-Devonian igneous activity: Large masses of peralkaline igneous rock, assigned by Billings (1956) to the White Mountain Plutonic Series and now dated as Jurassic by Lyons and Faul, (1968) extend into the southern part of the Dixville quadrangle (Hatch, 1963). The only other post-Devonian rocks in the area are a few thin and scattered lamprophyre dikes, probably related to the White Mountain Plutonic Series, in the Cupsuptic, Arnold Pond, Errol, and Dixville quadrangles.

Metamorphism

Regional metamorphism increases from chlorite grade, generally north of the 45th parallel, to upper sillimanite grade (Guidotti, 1966) in the south-central and southeastern part of the Oquossoc quadrangle. Green and Guidotti (1968, p. 265, fig. 19-5) show the garnet and staurolite isograds trending westward just south of the 45th parallel in the Oquossoc and northeastern part of the Errol quadrangles. From there, these isograds curve southwestward into northern New Hampshire. Devonian intrusive rocks are surrounded by sillimanite-grade rocks in the staurolite zone and by narrow cordierite-andalusite-bearing contact aureoles in the garnet zone (Green, 1963). In the chlorite zone, the slate adjacent to the Devonian intrusive rocks in the Cupsuptic area has been contact metamorphosed to cordierite-andalusite and cordierite-sillimanite hornfelses (Harwood, 1969b).

Green and Guidotti (1968, p. 265) report rotated porphyroblasts of garnet and staurolite in the regionally metamorphosed schist and infer that the metamorphism preceded or accompanied the latest Acadian deformation adjacent to the larger plutons. Radiometeric K/Ar studies of micas from the Errol quadrangle, however, indicate a Carboniferous age (Hurley and others, 1958) ascribed to a late Paleozoic overprint by Lyons and Faul (1968, p. 310). Slate in the chlorite zone does not show any evidence of ever having been at a higher metamorphic grade; therefore, it is concluded that the pre-Silurian rocks were regionally metamorphosed only to chlorite grade, if at all, during the Taconic orogeny.

Major unsolved problems

The Second Lake anticline (fig. 1) proposed by Harwood (1969a) has not been unequivocally established. Some or all of the rocks in its core could be Silurian or Devonian, or both, and thus form a homoclinal northwest-facing sequence at least as far northwest as the trace of the Frontenac syncline (fig. 1) as proposed by Green and Guidotti (1968), Green (1968), Hatch (1963), and Marleau (1968). Detailed investigations at the lithologic contacts and a diligent search for fossils, if successful, could establish or refute the Second Lake anticline.

A careful search for fossils in the black slate member and possibly the Magalloway Member of the Dixville might prove rewarding in the chlorite-grade rocks north and west of the Albee. Is the type Dixville Middle Ordovician, as concluded in this report and assumed by all previous investigators, or is it significantly older, say Middle Cambrian to Early Ordovician, and correlative with the Ottauquechee and lower part of the Stowe Formations?

References cited

Albee, A. L., 1961, Boundary Mountain anticlinorium, west-central Maine and northern New Hampshire: U. S. Geol. Survey Prof. Paper 424-C, p. C51-C54.

Berry, W.B.N., 1961, Graptolite fauna of the Poultney Slate: Am. Jour. Sci., v. 259, no. 3, p. 223-228.

Billings, M. P., 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. America Bull., v. 48, no. 4, p. 463-566.

_______________, 1956, The geology of New Hampshire, Part II--Bedrock geology: Concord, New Hampshire State Planning and Development Comm., 203 p.

Boucot, A.J., L961, Stratigraphy of the Moose River synclinorium, Maine: U. S. Geol. Survey Bull. 1111-E, p. 153-188.

Cady, W. M., 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull. v. 71, no. 5, p. 531-576.

Doll, C. G., Cady, W. M., Thompson, J. B., Jr., and Billings, M. P., Compilers and editors, 1961, Centennial geologic map of Vermont: Montpelier, Vt., Vermont Geol. Survey scale 1:250,000.

Green, J. C., 1963, High-level metamorphism of pelitic rocks in northern New Hampshire: Am. Mineralogist, v. 48, nos. 9-10, p. 991-1023.

_______________, 1964, Stratigraphy and structure of the Boundary Mountain anticlinorium in the Errol quadrangle, New Hampshire-Maine: Geol. Soc. America Spec. Paper 77, 78 p.

_______________, 1968, Geology of the Connecticut Lakes-Parmachenee area, New Hampshire and Maine: Geol. Soc. America Bull., v. 79, no. 11, p. 1601-1638.

Green, J.C., and Guidotti, C. V., 1968, The Boundary Mountains anticlinorium in northern New Hampshire and northwestern Maine: in Zen, E-an, and others, editors, Studies of Appalachian geology--northern and maritime: New York, Interscience, p. 255-266.

Guidotti, C. V., 1966, Variations of the basal spacings of muscovite in sillimanite-bearing pelitic schists of northwestern Maine: Am. Mineralogist v. 51, nos. 11-12, p. 1778-1786.

References cited (continued)

Harwood, D. S., 1966, Geology of the Cupsuptic quadrangle, Maine: U. S. Geol. Survey Open-file report, 259 p.

______________, 1968, Geology of the Arnold Pond quadrangle, Oxford and Franklin Counties, Maine: U. S. Geol. Survey open file report, 14 p.

______________, 1969a, The Second Lake anticline--a major structure on the northwest limb of the Boundary Mountain anticlinorium in northern New Hampshire, west-central Maine and adjacent Québec: U. S. Geol. Survey Prof. Paper 650-D, p. D106-D115.

______________, 1969b, Variations in the Δ-index of cordierite around the Cupsuptic pluton, west-central Maine: Am. Mineralogist, v. 54, p. 896-908.

Harwood, D. S., and Berry, W.B.N., 1967, Fossiliferous lower Paleozoic rocks in the Cupsuptic quadrangle, west-central Maine: U. S. Geol. Survey Prof. Paper 575-D, p. D16-D23.

Hatch, N. L., Jr., 1963, Geology of the Dixville quadrangle, New Hampshire: New Hampshire Dept. Resources and Econ. Devel., Bull. 1, 81 p.

Hitchcock, C. H., 1877, Geology of New Hampshire, Part 2, Stratigraphical geology: Concord, N.H., 685 p.

Hurley, P. M., Fairbairn, H. W., Pinson, W. H., Winchester, J. W., and Gheith, M. A., 1958, Variations in isotopic abundances of strontium, calcium, and argon and related topics: 6th Ann. Prog. Rept. to U. S. Atomic Commission, Massachusetts Inst. Technology, Dept. Geology and Geophysics, 193 p.

Logan, W. E., 1863, Report on the geology of Canada: Canada Geol. Survey Rept. Prog. to 1863, 983 p.

Lyons, J. B., and Faul, H., 1968, Isotope geochronology of the northern Appalachians, in Zen, E-an, and others, editors, Studies of Appalachian geology--northern and maritime: New York, Interscience, p. 305-318.

Marleau, R. A., 1957, Preliminary report on East Megantic and Armstrong areas, electoral districts of Frontenac and Beauce, Québec: Québec Dept. of Mines, Geol. Survey Branch Prelim. Rept. 336, 6 p.

______________, 1959, Age relations in the Lake Megantic Range, southern Québec: Geol. Assoc. Canada Proc., v. 11, p. 129-139.

References cited (continued)

Marleau, R. A., 1968, Woburn-East Megantic-Armstrong area, Frontenac and Beauce Counties: Québec Dept. Nat. Resources Geol. Rept 131, 55 p. (also French edition).

McGerrigle, H. W., 1935, Mount Megantic area, southern Québec, and its placer-gold deposits: Québec Bur. Mines Ann. Rept. 1934, pt. D, p. 63-104.

Moench, R. H., 1969, The Quimby and Greenvale Cove Formations in western Maine: U. S. Geol. Survey Bull. 1274-L, p. Li-L17.

Osberg, P. H., Moench, R. H., Warner, Jeffrey, 1968, Stratigraphy of the Merrimack synclinorium in west-central Maine in Zen, E-an and others, editors, Studies of Appalachian geology--northern and maritime: New York, Interscience, p. 241-253.

Theokritoff, George, 1964, Taconic stratigraphy in northern Washington County, New York: Geol. Soc. America Bull., v. 75, no. 3, p. 171-190.

U. S. Geological Survey, 1965, Geological Survey research 1965: U. S. Geol. Survey Prof. Paper 525-A, 376 p.

Zen, E-an, 1967, Time and space relationships of the Taconic allochthon and autochthon: Geol. Soc. America Spec. Paper 97, 107 p.

Road Log - trip A-3

Assemble at 8:00 A.M. at the junction of Routes 16 and 4 (at the Mobil Station) in the town of Rangeley. Stops will be in the Rangeley, Oquossoc, Cupsuptic, Errol, and Second Lake 15-minute quadrangles.

Mileage

0.0 Head east on Route 4 from assembly point through the town of Rangeley toward Farmington.

2.3 STOP 1: (Vehicles turn around and face west on Route 4). Large roadcuts in lower polymict conglomerate part of Rangeley Formation. Beds of pebble, cobble, and boulder conglomerate 3 to 20 feet thick grade into beds of feldspathic metasandstone 3 to 10 feet thick. Clasts vary from nearly spherical to tabular and are composed of finely laminated ("pinstripe"?) quartzite, vein quartz, coarse-grained gray granodiorite and quartz diorite (some containing blue quartz), fine-grained biotite granite, light-gray felsic metavolcanic rocks, black biotite-spotted lamprophyre and various dark-gray to black schistose biotite-rich rocks of uncertain origin. Note the blue quartz grains in the matrix of the conglomerate and in the interbeds of metasandstone. Look closely at the clasts in the conglomerate, noting particularly the granitic rocks and the laminated quartzite cobbles; future stops will be in possible source rocks for these clasts.

Walk west along the north side of Route 4 past small outcrop of massive, light-tan metasandstone, which Moench (1969) places at the base of the Rangeley Formation, to larger roadcut in light- and dark-green to gray-green laminated metasiltstone and metashale. This rock is part of the Greenvale Cove Formation assigned to the Late Ordovician (?) by Moench (1969). South of Rangeley Lake, the Greenvale Cove contains interbeds of calc-silicate rock and biotite granofels. Drive west on Route 4 toward Rangeley. Magnificent view of Rangeley Lake and Bald Mountain to west-southwest and Spotted Mountain and Ephraim Ridge to west-northwest.

4.6 Town of Rangeley - pass assembly point and continue west on Routes 16 and 4 toward the village of Oquossoc.

5.8 Roadcuts on top of hill west of Rangeley and scattered along both sides of Routes 16 and 4 for the next 2 miles are in the upper metashale member of the Quimby Formation (Moench, 1969).

8.4 STOP 2: Lower metagraywacke member of Quimby Formation consisting of 1/2 to 3-foot beds of light-gray to brown, light rusty tan-weathering, fine- to coarse-grained metagraywacke. There is a bed of conglomeratic metagraywacke at the west end

of the roadcut (NO HAMMERS PLEASE; the clasts show up best on the weathered surface). The intensely stretched pebbles are composed of vein quartz, felsic metavolcanic rocks, and gray quartzite.

Beds dip gently to the southeast; graded bedding indicates both upsidedown and right-side-up beds. The outcrop is near the hinge line of a north-northwest-trending syncline superimposed on earlier possible slump structures. The Quimby is gradationally above black slate assigned to the Dixville Formation, which to the north contains Middle Ordovician (zone 12) graptolites.

Proceed west on Routes 16 and 4.

10.2 Enter Oquossoc quadrangle.

10.3 STOP 3: Low roadcut in black slate of Dixville Formation continuous with graptolite-bearing slate to north in the Cupsuptic quadrangle. Note that the northwest-trending cleavage is crenulated by a later west-trending, north-dipping fracture cleavage. Walk uphill into woods and follow slope to southwest parallel to the road past several more low outcrops of black slate to low hogback ridge of black slate breccia cemented with white vein quartz. Outcrops southwest of the hogback ridge are greenstone assigned to the metavolcanic member of the Dixville. The contact here is a northwest-trending fault with apparent northeast-side-down movement. On strike to the northwest the fault crosses into the greenstone, and the greenstone-black slate contact is marked by interfingering lenses of the two lithologies. The greenstone contains massive, foliated, and agglomeratic layers intermixed in variable proportions. It is composed of actinolite, albite, chlorite, quartz, calcite, epidote, sphene, magnetite and pyrite typical of the greenschist facies of regional metamorphism.

Proceed southwest on Routes 4 and 16.

11.1 Junction of Routes 16 and 4; continue west on Route 4 through the village of Oquossoc.

12.3 Turn right (north) onto dirt road just east of yellow wooden garage and proceed northwest toward Spots Point.

12.7 STOP 4: (Vehicles turn around). Outcrops on shore of Mooselookmeguntic Lake at Spots Point. Green phyllite containing abundant pods and stringers of quartz typical of the Deer Mountain Member of the Albee Formation of Green and

Guidotti (1968). Quartz stringers are commonly crinkled and folded about the north-northwest-trending cleavage, producing a distinct lineation that plunges N.35°E. Beds of greenstone 2 to 6 feet thick are scattered in the green phyllite and become more abundant near the contact with the overlying metavolcanic member of the Dixville.

The Deer Mountain Member differs from the typical Albee in its abundance of quartz stringers and its general lack of pinstripe arenaceous beds. In this respect, the Deer Mountain Member is lithologically similar to the noncarbonaceous part of the Aziscohos Formation, but the Deer Mountain Member is near the top of the Albee, and the Aziscohos is below the typical Albee.

Retrace the route east through Oquossoc to the junction of Routes 16 and 4.

14.3 Turn left (northwest) onto Route 16 from Route 4 at the Rangeley Region Chamber of Commerce information center. Several new roadcuts in the metavolcanic member of the Dixville.

15.8 Cross Kennebago River.

17.3 Enter Cupsuptic quadrangle at the south end of Cloutman Ridge.

19.8 Cross Cupsuptic River.

21.4 Enter Oquossoc quadrangle on the west side of Cupsuptic Lake; proceed south up steep hill on Route 16.

23.1 STOP 5: Coarse-grained, gray and pink, strongly sheared and metamorphosed Adamstown Granite of Green and Guidotti (1968) composed of quartz, microcline, plagioclase, biotite and minor muscovite. The pink alkali feldspar is not present throughout the pluton and the intensely sheared texture is prominent only in the northern part. In the central part of the pluton the granite is less sheared and more thoroughly recrystallized into a coarse-grained, equigranular to porphyritic metagranite.

The Adamstown Granite intrudes the Albee and Aziscohos Formations along its north and east sides and is intruded and metamorphosed by a large body of Devonian adamellite on its south and west sides. It is clearly older than the Devonian intrusive rocks and is assigned to the Highlandcroft Plutonic Series of Billings (1956) of Ordovician age.

This granite could be one possible source for the granitic clasts in the polymict conglomerate of the Rangeley Formation. It must be remembered, however, that the pluton was more intensely metamorphosed during the Devonian Acadian orogeny than was the conglomerate we saw at STOP 1, and any comparison must acknowledge this complicating factor.

Proceed southwest on Route 16.

27.9 Richardson Pond north of the road. Some new roadcuts in the gray facies of the Adamstown Granite.

31.1 New roadcuts of gray Adamstown Granite containing inclusions and septa of the Aziscohos Formation (?).

31.8 STOP 6: Low outcrop on the south side of Route 16 of light-gray biotite-muscovite adamellite assigned to the New Hampshire Plutonic Series of Billings (1956). The granite is composed of quartz, oligoclase, microcline, muscovite, biotite, rare garnets, and common accessories and may be a product of Acadian anatexis.

Proceed northwest on Route 16 toward Aziscohos Dam.

32.5 Lunch STOP: Dam at south end of Aziscohos Lake. This dam, built in 1911, flooded about 13 miles of the Magalloway River valley. Foothills and summit of Deer Mountain underlain by Albee Formation visible across lake to northeast. The dam is on or very near the contact between the 2-mica adamellite seen at the last stop and the carbonaceous lower part of the Aziscohos Formation (see large blocks of rusty-weathering schist near boat launching area). Rocks here are in the sillimanite zone.

32.6 Enter Errol quadrangle--proceed west toward Wilsons Mills.

34.2 Turn right (north) at roadside cafe onto private Brown Co. logging road to Parmachenee Lake.

34.7 STOP 7: Park on side of dirt road; walk east about 100 feet to outcrops in Abbott Brook. Light-gray to gray-green schist with abundant stringers and pods of quartz typical of the upper, noncarbonaceous part of the Aziscohos formation. Quartz stringers, color laminations, and a few scattered feldspathic siltstone beds trend northwest and dip steeply or are vertical. Several early isoclinal minor folds with northwest-trending axial surfaces and gently to moderately steep plunging axes (generally plunging to southeast) are shown by the quartz stringers. These northwest-trending features are clearly refolded by later isoclinal folds that trend N.40°E. and plunge steeply to the northeast. A

dominant northeast cleavage parallels the axial surface of the late folds, but there is no pervasive northwest-trending cleavage associated with the early folds. This structural style of late northeast folds superimposed on early northwest folds is common in the pre-Silurian rocks in the Cupsuptic quadrangle and is invoked on a regional scale to explain the pattern of the Albee-Dixville contact (see figure 1).

The schist here is in the staurolite zone and contains quartz, plagioclase, muscovite, biotite, garnet, staurolite, late chlorite, and accessories. Magnetite porphyroblasts are common in some zones, and andalusite is locally present with staurolite, garnet, and biotite.

38.4 Enter Second Lake quadrangle.

Continue north on Parmachenee road west of Aziscohos Lake.

43.1 STOP 8: Pavement outcrops on west side of road consisting of green phyllite and interbedded light-gray finely laminated "pinstripe" quartzite characteristic of the bulk of the Albee Formation in the chlorite zone. The abundance of arenaceous beds with their diagnostic micaceous laminae and the marked decrease in the number of quartz stringers serve to distinguish the Albee from the noncarbonaceous Aziscohos which is gradationally below it. Note that the light- and dark-green color banding in much of the phyllite is not parallel to the trend of the arenaceous beds. This color banding may not have been a primary sedimentary feature or, if it was, the present disparity in trends may reflect differential flowage between the arenaceous and pelitic beds during deformation. Steep N.40°E. axial-plane cleavage dominates, with many steeply-plunging folds giving a general N.70°E. trend to the bedding.

49.5 Cross West Branch of Magalloway River near Long Pond; continue east.

50.2 Turn south at cross road marked by Parmachenee Club sign.

50.6 Enter Cupsuptic quadrangle.

51.6 Cross Magalloway River at the head of Aziscohos Lake.

51.7 Turn left (north) at cross road; proceed north on east side of Magalloway River gorge.

53.9 Continue north on main haulage road toward east side of Parmachenee Lake.

54.6 STOP 9: Pavement outcrop of gray-weathering black to silvery-gray phyllite of the Dixville Formation. Bedding is not distinct here, and the main planar feature is a northeast-trending essentially vertical cleavage. On the slope to the east black slate and tan-weathering feldspathic metasandstone are interbedded in variable proportions. The beds trend generally southeast and dip steeply east.

This belt of black slate, which apparently is continuous with the type Dixville at Dixville Notch, New Hampshire (See Green, 1968) has not been unequivocally dated as yet. It contains Protospongia sp. of Cambrian or Ordovician age at locality 0-3 (fig. 1) about 5 miles east of here and is correlated, in part, with the graptolite-bearing black slate at locality 0-1 on the basis of lithologic similarity and stratigraphic position above the Albee.

55.6 Cross Moose Brook.

55.8 Turn right onto logging road that crosses the low hills immediately east of the Parmachenee farm.

56.9 STOP 10: Park in wood yard and walk about 500 feet north on abandoned logging road. We will look at several outcrops in this immediate area that are in small synclinal outlier of Silurian rocks surrounded on three sides at least by black slate of the Dixville Formation.

The lowest unit on the north limb of the syncline is light- and dark-gray pebble to cobble conglomerate containing interbeds of feldspathic quartzite, gray metasiltstone, and minor purplish-red slate. The clasts of the conglomerate are predominantly vein quartz, chert, and gray quartzite, but less stable fragments of black slate, feldspathic siltstone, and some green slate are locally present near the base of the unit. A discontinuous layer of polymict conglomerate about 30 feet thick is present on the south limb of the inlier. Throughout much of the conglomerate the pebbles are nearly spherical, but here they tend to be elongated down the dip of bedding.

The conglomerate is separated from the overlying tan-weathering argillaceous limestone by an east-northeast-trending normal fault. Shearing adjacent to the fault has obliterated most of the bedding in the argillaceous limestone and apparently has stretched the pebbles in the conglomerate. About 100 feet from the fault an outcrop of typical argillaceous limestone clearly shows the bedding characteristics and the more typical, relatively open gently plunging folds characteristic of the Acadian deformation in the Silurian rocks of this area.

The argillaceous limestone is overlain by well-bedded pit-weathering fossiliferous arenaceous limestone. Excellent exposures are available about 200 feet south of the wood yard. Bedding in the arenaceous limestone strikes east and dips south. To the west near the east shore of Parmachenee Lake the arenaceous limestone is in contact with the conglomerate along a normal fault that is essentially parallel to the one separating the conglomerate and argillaceous limestone. Thus the calcareous rocks are interpreted to be in a narrow northeast-trending graben bounded to the north and at least partly on the south by the basal conglomerate unit. Elsewhere on the south limb of the synclinal inlier, the basal unit is composed of gray argillite and tuff (?) similar to material interbedded in the conglomerate on the north limb. This is the last stop on this trip; return to Rangeley.

Pre-Silurian Flysch Relict Structures in Cordierite-K-Feldspar Granofelses, Long Falls of Dead River, Somerset County, Western Maine

Gary M. Boone
Syracuse University

Introduction

The youngest pre-Silurian stratigraphic unit in the Little Bigelow Mountain and Pierce Pond quadrangles consists of Mg-rich metapelite and metagraywacke.[1] The unit is informally designated here as the Dead River Formation. It has been subjected to a wide range of metamorphism, from chlorite-grade to the north and east, to K-feldspar-cordierite grade at Long Falls and elsewhere to the west, along the north margin of the Flagstaff igneous complex (Fig. 1). At chlorite grade the metapelite is characteristically green to silvery greenish gray, and commonly pin-striped with quartz-rich laminae parallel to cleavage and bedding. Axial plane cleavage and, in some examples, a weak mimetic bedding-plane schistosity generally obscure primary depositional structures in the low-grade chlorite-rich phyllites. Where the formation is wedged between the Pierce Pond gabbro and Lexington batholith, recrystallization at higher metamorphic ranks has partly destroyed the cleavage and enhanced the visual contrast between pelitic beds, laminae, and lithic fragments of slightly different composition. Along the strike of the Dead River Formation in this structural wedge, the metapelitic interbeds of the formation range from schistose hornfelses with weakly developed mimetic bedding schistosity in the eastern part of the wedge, to gneissose granofelses, to the west, where again, the geneissosity appears to be a dominantly mimetic feature. With local exceptions, the parts of the formation that have been subjected to medium- to high-grades of metamorphism have experienced very slight penetrative deformation. Because of logistic constraints, we will confine our observations to the upper part of the Dead River Formation in the vicinity of Long Falls. Fortunately, despite (or more likely because of) the high metamorphic grade, relict depositional features are exceptionally well preserved here, and the stream-worn surfaces in the gorge of Long Falls provide the control in three dimensions for adequate structural interpretation for detailed, as well as large scale features.

Acknowledgements

The mapping was carried out as part of an ongoing program supported by the Maine Geological Survey. I thank the many colleagues,

---

[1] The Dead River Formation is traced along regional strike southwestward into, and is correlated with the Albee Formation (see Boone and others, this volume, p. 12).

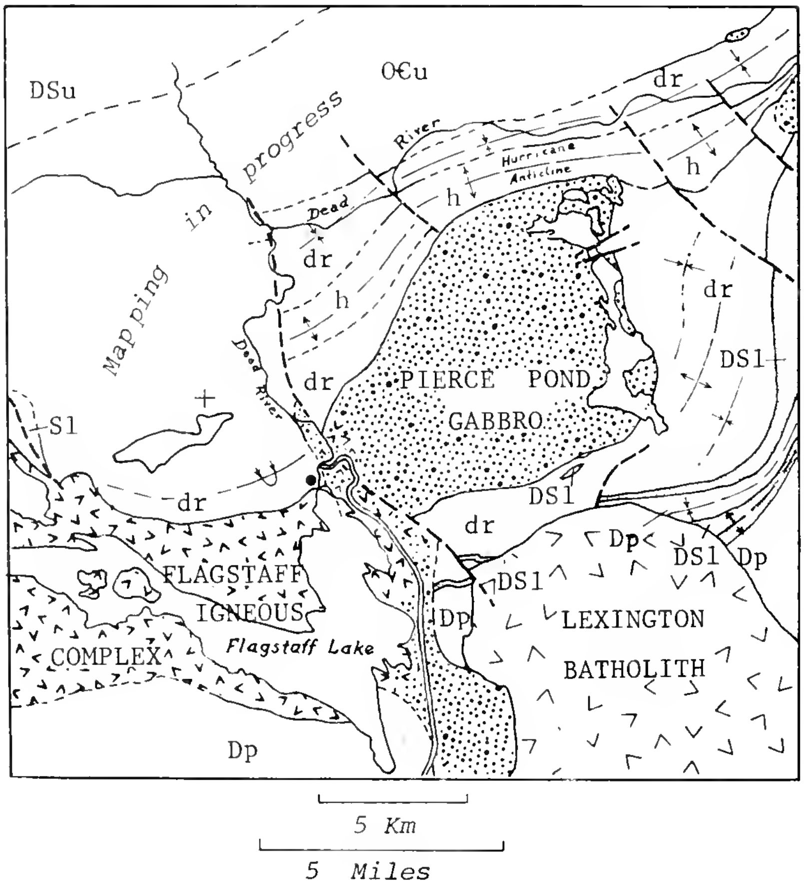


Fig. 1. Generalized geologic map of eastern part of Dead River drainage basin, southeastern Spencer and southern Pierce Pond quadrangles, northeastern Stratton (A. Griscom, unpublished data), and northern Little Bigelow Mountain quadrangles. (+ : quadrangle corners.) Paved access road shown near east shore of Flagstaff Lake; area of detailed map (Fig. 2) shown by dot at end of road. Two youngest pre-Silurian formations, from older to younger: h = Hurricane Formation; dr = Dead River Formation (informal designations). Siluro-Devonian units: p = metapelite; l = limestone and calc-silicate rock; u = undifferentiated metasedimentary units.

too numerous to list here, whom I have subjected to discussion of these features. Dr. R.B. Neuman, U.S. Geological Survey, kindly examined calc-silicate rock thought to contain a fossiliferous shelly fauna. I also thank Mr. and Mrs. J. Arthur Haskell for their help and cooperation during my mapping in this area.

Sequence of Metamorphic and Igneous Events

These events are summarized before discussing the sedimentology, in order to document and discriminate the later deformational and metamorphic features from those that are primary and depositional.

The trend of major folds in the vicinity of Long Falls is east-northeasterly, parallel to the contact zone of the Flagstaff igneous complex, but normal to the south west margin of the Pierce Pond gabbro (Fig. 1). The major folds are refolded locally (Fig. 2). They appear unsystematic with regard to orientation of the secondary axial surfaces, but the limbs tend to conform to local trends of intrusive contacts. Intrusion-tectonics thus appears to be responsible for the secondary open folding shown on the detailed map. The major folds are overturned to the northwest and in this respect, they reflect the regional structure in the pre-Silurian beyond the thermal aureoles. Plunge of major folds here is variably gentle, probably averaging nearly zero. The final stage of folding involved flexural slip between layers of different competency; this produced locally strong 'a' lineation marked by small-scale grooves in quartz-feldspar-rich surfaces of the granofelses. These microcorrugations are not to be confused with relict flute- and groove-casts preserved in one section of the gorge. The metamorphic fabric is intimately involved and expresses the lineation; it is not similarly involved in the preservation of the flute- and groove-casts. There will be a discussion at one locality of the question of fossils acting as possible strain-gauges in an interbed of calc-silicate rock.

Large, gently southward-dipping sheets of fine-grained diabasic gabbro represent the first stage of igneous activity. They appear sill-like in plan, and are approximately concordant to the orientation of axial surfaces of the major folds, but are discordant to fold limbs (see composite cross-section, Fig. 2). Where their contacts against the granofelses of the Dead River Formation are unmodified by later intrusion of quartz monzonite, the contacts are microscopically sharp. No evidence of assimilation or metasomatic alteration is evident, nor are any xenoliths of the granofels to be found in the mafic sheets. Despite abundant cordierite in the granofels, none has been detected in stained sections of the gabbro at the contact. This in fact is expectable in terms of the thermal divide on the liquidus between normal gabbroic, and mafic aluminous rock compositions discussed by Chinner and Schairer (1962), which would prevent mixing by

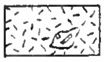
Figure 2 Explanation

Igneous and Metamorphic Rocks

Post-Acadian

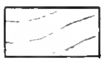
Medium- to coarse-grained, porphyritic quartz monzonite of Flagstaff Igneous Complex. Inclusions of metagabbro shown schematically where abundant.

Late Acadian



Fine-grained, biotite-hornblende-bearing metagabbro of Flagstaff Igneous Complex. Apophyses of quartz monzonite shown schematically where abundant.

Lower Ordovician(?)

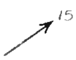


Dead River Formation (informal designation); upper part. Cordierite-biotite-K-feldspar-plagioclase-quartz granofels and metagraywacke; subordinate zoisite-bearing, biotitic quartzite. Thin- to medium-bedded relict graded sets in lower parts of local section. Relict graded, wavy, and flaser bedding with veins, dikes and sills of metagraywacke common. Local sequences of climbing-ripple sets.

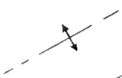
Structure Symbols

Strike and dip of normally facing beds

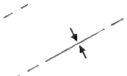
Strike and dip of overturned beds



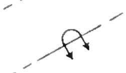
Bearing and plunge of fold axes



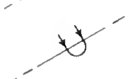
Anticline



Syncline



Overturned anticline



Overturned syncline

Crestal trace
Trough surface trace

Fig. 2. Geologic map of Long Falls, Little Bigelow Mountain quadrangle. Structural relations projected into N - S section.

assimilation. Lack of assimilation and bimodal separation of quartz norite and cordierite-rich aluminous rock compositions in the eastern Scottish Highlands (Gribble and O'Hara, 1967) also support the prediction from the experimental phase relations.

The gabbro here is spatially related to the border zone of the Flagstaff complex, but is probably part of the regional emplacement of mafic magmas as exemplified in part by the Flagstaff composite intrusion and the Pierce Pond gabbroic stock. Radiometric age determinations on mafic and other igneous rocks are cited in the introductory article; structural relations at contacts of mafic intrusives (Boone, this guidebook, trip A-2) indicate pre-latest Acadian emplacement.

Except for the central parts of the thickest dike-like bodies, the gabbros are partly altered, containing the excess number of phases: augite, hornblende, biotite, magnetite, ilmenite, hypersthene(?), quartz (rare), and zoned andesine-labradorite replaced in checkerboard pattern by more sodic plagioclase. Sparse patches of quartz enclosing cummingtonite and biotite have been found.

The emplacement of porphyritic quartz monzonite was responsible for much of the metamorphism of the Dead River Formation, and of the gabbroic sheets as noted above. Were it not for the extensive retrograding of the gabbro, and uniformity of prograde metamorphism in the Dead River Formation, assigning most of the metamorphism to 'granite' emplacement would not be warranted. Implied here is a much greater mass of subjacent 'granite' than is represented by the sub-concordant, dike-like sheets, and discordant pipes and apophyses of quartz monzonite at the present level of erosion. The alternative implication of a large subjacent mass of mafic rock is discounted on the grounds that the K-feldspar-cordierite zone is mapped on the scale of Figure 1 as spatially associated with the northern quartz monzonite unit of the Flagstaff complex.

Figure 2 shows that the quartz monzonite intruded along principal zones of weakness marked by gabbro-Dead River Formation contacts. These zones apparently created a similarity in structural style of emplacement, mimicking that of the gabbro sheets. Although slight chilling is evident in the relict texture of marginal gabbro, no decrease of grain-size is evident in quartz monzonite at the margins of the quartz monzonite sheets. Prominent reaction zones of biotite developed where quartz monzonite intruded gabbro. The biotite is clearly a product of reaction of potassic, hydrous magma with mafic components of the gabbro. A simple equation representing a hypothetical reaction that would not create excess alumina in the quartz monzonite liquid, is:

$$KALSi_3O_{8(l)} + H_2O_{(l)} + 3FMSiO_{3(s)} \rightarrow K(FM)_3AlSi_3O_{10}(OH)_{2(s)}. \quad (1)$$

components in qtz-monzonite liquid	hypersthene in gabbro	biotite reaction zone at contact

Biotite is the sole mafic silicate that was stable with liquid at the time of quartz monzonite emplacement. Sparse almandite-rich, Mn-bearing garnet crystallized in the quartz monzonite in contact with wall rock and inclusions of Dead River granofels.

Metamorphism of the Dead River Formation

The granofels at Long Falls represents the highest rank in a continuous metamorphic progression across the strike of the formation that has been documented from muscovite-chlorite grade. Bulk chemical compositions of the aluminous layers across isograds are very similar, and show no detectable systematic variation. This fact is important in the discussion below, regarding the question of anatexis and the origin of the quartzofeldspathic veins in the granofels.

The granofels is a 'clean'-textured, equigranular, slightly gneissic rock containing the quartz-bearing equilibrium assemblage:

cordierite (Mg/Mg + Fe + Mn = 0.62, molar ratio)
biotite (Ti-rich, Mg/Mg + Fe ≈ 0.45)
K-feldspar (slightly perthitic; Or/Or + Ab + An ≈ 70 mole%)
plagioclase (slightly zoned; An/Or + Ab + An ≈ 35 mole%)

in addition to minor magnetite and traces of retrograde muscovite in plagioclase. Proportions of phases vary across thin laminae and thin, subtly-graded beds ranging from a millimeter to a few centimeters thick.

A bed of sulfidic, calc-silicate rock bearing probable remains of a shelly fauna, contains the following mineral assemblages, listed in order of decreasing abundance:

white lithic fragments		green matrix
plagioclase (An_{60-70})		quartz
clinozoisite		ferrosalite
salite		zoisite
calcite		plagioclase (An_{60})
quartz		Fe-sulfide
sphene	(minor)	
tremolite		

Assuming the prograde assemblages in the mafic-aluminous granofels and calc-silicate rock were formed in equilibrium with the retrograde assemblage in the gabbro sheets, and with the phases crystallizing from the quartz monzonite melt, the isograde assemblages in these varied bulk compositions in the K-feldspar-cordierite zone at Long Falls are

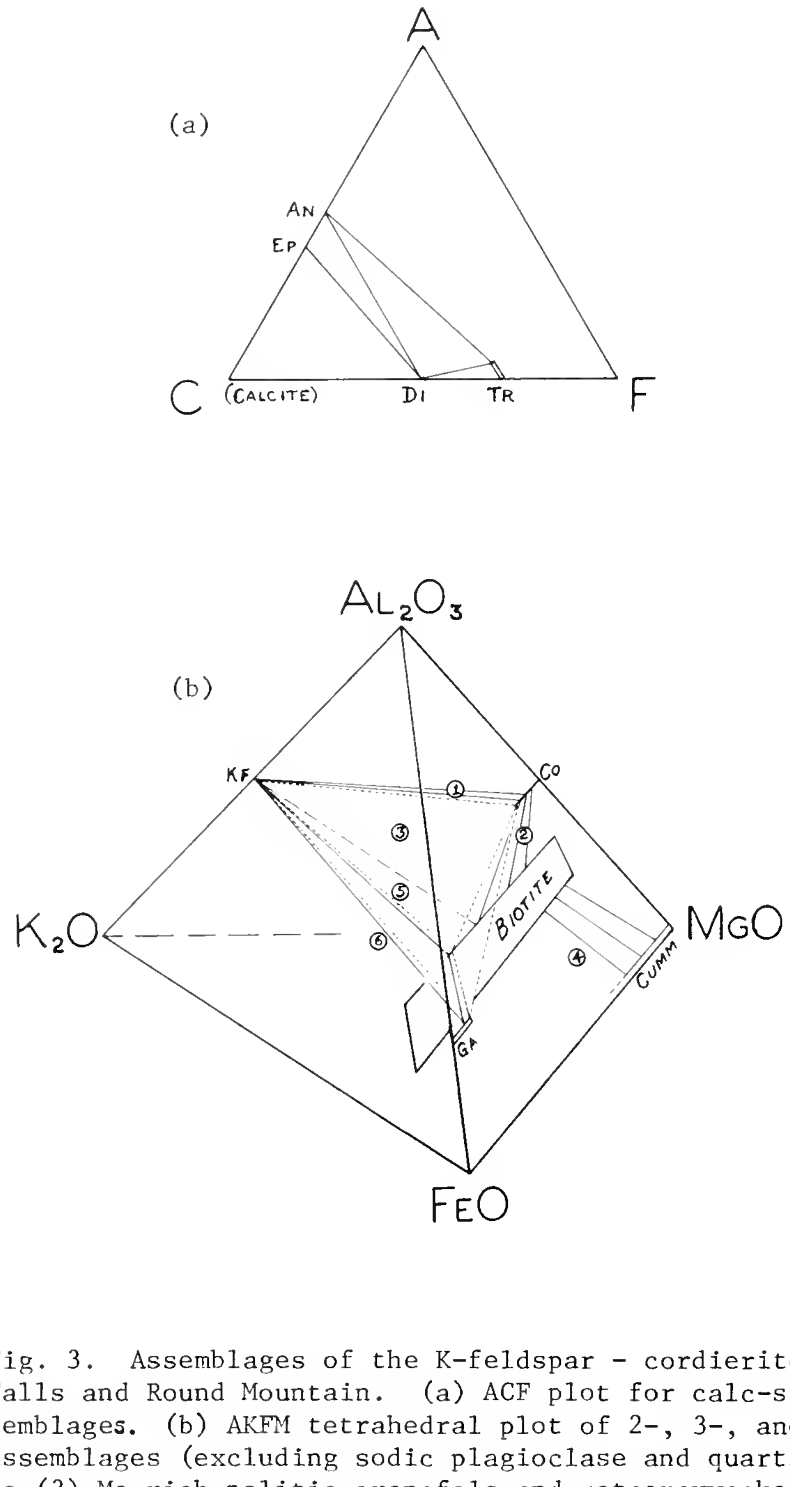


Fig. 3. Assemblages of the K-feldspar - cordierite zone, Long Falls and Round Mountain. (a) ACF plot for calc-silicate assemblages. (b) AKFM tetrahedral plot of 2-, 3-, and 4-phase assemblages (excluding sodic plagioclase and quartz) for: (1) to (3) Mg-rich pelitic granofels and metagraywacke; (4) cummingtonite-biotite-quartz patches in metagabbro; (5) K-feldspar-cordierite-almandite-biotite granofels (dotted lines), Round Mountain; and (6) almandite-rich garnet in quartz monzonite in contact with wall-rock and inclusions of pelitic granofels.

shown in Figure 3. Minor calcite persists with quartz in the calc-silicate system in which all alumina is fixed in associated phases; this suggests that CO_2 did not readily escape from the central parts of the thickest (10 cm) layers. Wollastonite has not been found.

Pelitic compositions sufficiently rich in Fe^{2+} to form the assemblage biotite + cordierite + almandite + K-feldspar have not been found at Long Falls, but this assemblage has been documented in Fe-rich, pelitic granofels three miles to the south, and is included in Fig. 3.

Muscovite is a characteristic mineral in the formation at lower metamorphic grades and persists throughout the sillimanite zone in the structural wedge of the formation east of Long Falls. At K-feldspar-cordierite grade, muscovite was probably consumed in reactions summarized by the following equation:

$$\text{biotite}_a + \text{muscovite} + \text{sillimanite} + \text{quartz} + \text{magnetite} + \text{hydrogen} \longrightarrow \text{biotite}_b + \text{K-feldspar} + \text{cordierite} + H_2O.$$

$$KFe_{1.5}Mg_{1.5}Al_{1.5}Si_{2.5}O_{10}(OH)_2 + KAl_3Si_3O_{10}(OH)_2 + Al_2SiO_5 + 4SiO_2 + 2/3Fe_3O_4 + 2/3H_2 \longrightarrow$$

$$KFe_{2.83}Mg_{0.17}Al_{1.5}Si_{2.5}O_{10}(OH)_2 + KAlSi_3O_8 + Mg_{1.33}Fe_{0.67}Al_4Si_5O_{18} + 1\ 2/3\ H_2O. \qquad (2)$$

The equation approximates the composition of cordierite at Long Falls, and reflects the direction of change of Fe/Mg in biotite, as well as the modal relationship of biotite and magnetite in granofelses at muscovite-sillimanite and K-feldspar-cordierite grades.

Pressure-temperature conditions of metamorphism are deduced to have been in the range of 650° to 700°C at total pressure less than 4, but greater than ~2.5 kb (see location of highest P-T plot (Boone, trip A-2, Fig. 4) in which disappearance of muscovite occurs within the stability field of sillimanite, but at pressures within the stability range of biotite + cordierite + almandite).

Pinstripe laminae and cross-cutting veins contain the same minerals as do the surrounding pelitic layers, but they are much richer in plagioclase and quartz. A bimodal spectrum of thickness ranges from laminae to thick beds. Veins commonly thicken where they mutually intersect. The veins and laminae contain small, somewhat variable proportions of cordierite and biotite, and K-feldspar is invariably sparse. Plagioclase compositions in different laminae and veins vary from An_{18} to An_{31} mole percent.

The structural and metamorphic environment exhibits most, if not all the prerequisites for a high-level, low pressure anatectic paragenesis of the quartzofeldspathic laminae and veins. The

locally lower An-content of plagioclase, as compared to most of the pelitic layers, would favor this interpretation. With regard to compositions of low-melting granitic fractions, however, the anatectic model fails. Although K-feldspar (Or_{50-70}) is present, the proportion of plagioclase is 90-95 percent of total feldspar, well outside the range of plagioclase/K-feldspar ratios in low melting anatectic fractions (see Winkler's summary of experimental anatexis and studies of corresponding natural systems (1967, Chap. 16)). In other words, the trondhjemitic compositions of the laminae and veins (excepting cordierite) could be produced anatectically only at temperatures much above the granite solidus, in which case we would not expect to find K-feldspar, indeed high proportions of it, in the crystalline residue surrounding the veins. Minor concentrations of biotite line the surfaces of some of the laminae parallel to relict bedding, and indicate that at least some metamorphic differentiation has accentuated compositional differences across the relict bedding.

Some may wish to argue that the rocks are injection-migmatites by proposing a trondhjemitic, plagioclase-rich melt infiltrating the granofels. The cordierite and biotite in the trondhjemitic veins would require special explanation, as no larger masses of rock of such composition have been found in the region. Evidence to the contrary, however, lies in the similar chemical compositions of pelitic layers between quartz- and plagioclase-rich laminae and veins in rocks of the formation throughout the entire range of metamorphism. Thin sills and cross-cutting dikes of typically porphyritic Flagstaff quartz monzonite in the granofels at Long Falls appear to silently settle this issue.

The structural geometry and compositions that the laminae and veins display are more fully and simply explained by processes in the depositional environment. Therefore, we may now consider the primary origin of these and other features in the Dead River Formation.

Deposition of the upper part of the Dead River Formation

The Dead River Formation is approximately 2500 feet (~760 meters) thick, of which roughly 200 meters of interbedded metapelite and metagraywacke constitute the upper part.

The characteristically thin to medium, parallel beds of the lower and middle parts of the formation become progressively interrupted in the upper part by zones, up to 20 meters thick, of wavy bedding, climbing-ripple sets, and flaser bedding in chaotically-deposited portions. Small-scale slumps overlain by undisturbed beds also become commoner. Pull-apart structure, and convolute laminae possibly marking flow-rolls, are exposed in thicker beds in the upper part of the gorge at Long Falls.

Graded beds occur in many parts of the formation from the base to the highest exposed part (where it is disconformably overlain by Siluro-Devonian calc-silicate rock, or where intruded by the Flagstaff Igneous Complex). The grading is commonly difficult to detect, however, owing to slight compositional change across graded sets, or to thin bedding complicated by shearing. The latter problem is particularly met with in the lower metamorphic zones. Chips and lumps of pelitic rock fragments are common in beds of metagraywacke, and sparse occurrences of quartzwacke blocks, up to a few meters in length, in pelitic host rock (argille à blocs) have been noted west of Long Falls. Sole-markings at the base of quartz-rich beds have rarely been observed, probably because of the interplay of rock structure and erosion. Well-preserved examples, however, are exposed in the 'calcareous sands' (calc-silicate-bearing, biotitic quartzite) in the upper (southern) part of Long Falls.

Sufficient description has been introduced here to warrant turning to flysch sedimentology to account for many, if not all the relict structures and compositional assemblages. No attempt is made to review or redefine the meaning of flysch. The term is used here to connote a tectonic-sedimentologic synthesis. In brief, the entire formation, and that part of it in particular exposed at Long Falls, meet all twelve of the criteria (both inclusive and exclusive) of the diagnostic features of flysch discussed by Dzulynski and Walton (1965). The protolith of rock here termed metagraywacke was likely feldspathic, chlorite-rich sandstone that contained variable amounts of potassic micas. In the chlorite zone, graywacke interbeds in the formation contain roughly a percent of detrital K-feldspar (coarsely twinned microcline). The chlorite-rich matrix is matched at higher grades of metamorphism by equivalent proportions of cordierite and biotite. A relatively small proportion of the metagraywacke beds are calcareous (calc-silicate-bearing above the biotite isograd), but calcium-rich beds and lenses are rare. The chemical composition of the mafic aluminous metapelite that constitutes most of the formation is discussed below in regard to provenance and paleotectonic setting.

Where we enter the gorge, the flysch sequence is exposed in several overturned folds. Sections approximately parallel and transverse to paleocurrent directions can be studied. Parallel and wavy bedding predominate in this part of the gorge, but on the east side, near the old log-work crib overlooking the channel, excellent examples of sets of climbing ripples are exposed in an overturned section. At this locality linguoid ripples of two sets indicate current sources in the approximate range of N40°W - N70°W, after structural restoration. Sole markings (flute casts) at the base of overturned, dark gray biotitic quartzite beds in the upper part of the falls indicate current source from the SSE.

Some of the climbing-ripple sets in this part of the Dead River Formation are similar in form and scale to those produced by McKee (1965) in wave-tank experiments; others, however, are of much larger wavelength and stacked in thicker sets. Ripple-drift with sediment added from suspension is favored for the origin of these climbing-ripple sets. In the exposure mentioned above, climbing-ripple sets merge into parallel beds. The low angle between the ripple foresets and the planes bounding the sets suggests that current velocities in the lower flow regime were occasionally high (McKee, ibid., p. 80).

Many of the depositional sequences duplicate the truncated base cut-out sequences described by Bouma (1962), and enhance the hypothesis that the upper part, if not most of the Dead River Formation was deposited by repetitive influx of sediment from turbidity currents. Slumping and small-scale mass movement of semi-coherent sediment is also recorded in the upper 100 to 200 meters of the formation. Here, the abundance of 'sand' (metagraywacke) dikes and sills cutting across, and connecting parallel or wavy interbeds of metagraywacke lends a migmatitic appearance to the rocks. The 'sand' dikes commonly intersect each other in intervening beds of metapelite (granofels). No major compositional differences can be detected between veins, dikes, and the interbeds they connect. The dikes and veins are on the scale of "minor features" as described by Dzulynski and Walton (1965, p. 162). Many of the dikes and veins are deformed into simple curvilinear, or complex ptygmatic shapes. Because of the compositional similarity to the more extensive, parallel beds of metagraywacke, and because the axes of the contorted veins and dikes are not systematically related to the primary or secondary folds, their origin is assigned to the depositional environment of relatively deep-water flysch. The structures of many of the veins superficially resemble similar features described by Smith (1968) in the Belt-Purcell succession, which he attributes to a shallow-water organic origin. The forms also resemble those of veins in metamorphic parageneses, as described by Shelley (1968). Veins that are unquestionably of metamorphic origin, however, commonly are compositionally different than the host layers or bounding interbeds.

Convolute lamination is not abundant, nor, as Dott and Howard (1961) point out, is it unique to flysch, or to turbidity-current-deposited sediment. Many small-scale, 'chaotic' folds in restricted sections, however, resemble similar structures associated with deeply load-casted flutes (Kuenen, 1957, fig. 16) in turbidity current deposits.

Lastly, we may inquire if the Dead River Formation is characterized by redeposited material from turbulent suspensions associated with gravity-propelled density currents, and whether much of its detritus may first have accumulated on shelves or upper portions of continental or island rises. Aside from shale chips, occasional sandstone blocks, and pervasive, relict poor sorting in graywacke beds in the chlorite zone, two features deserve mention. The first is local, the second is formation-wide.

Table 1. Weight-proportions of critical oxides in shale - basalt mix required to approximate the composition of pelitic granofels, Long Falls

	Average Paleozoic shale[1]		Mg-rich metapelite, Long Falls	Basalt glass, Volcanics, Hound Island Alaska[2]	
SiO_2	63.7	0.8	60.48	0.2	47.95
TiO_2			1.28		1.75
Al_2O_3	17.5	0.4	17.05	0.6	16.91
FeO*	7.0	0.5	8.69	0.5	9.95
MnO	tr		0.21		0.17
MgO	2.5	0.8	4.22	0.2	7.04
CaO	1.5	0.95	1.83	0.05	11.09
Na_2O	1.1		2.00		2.47
K_2O	3.8		3.02		0.52
H_2O+			1.76		1.58
			100.54		

[1]Clarke's average; from Pettihohn, 1949, Table 61, p. 344.

[2]Muffler and others, 1969, Table 1, p. 198.

*Total iron expressed as FeO.

A thin, boudinaged bed of dark green calc-silicate rock in the gorge at Long Falls contains white calc-silicate 'calcareous mud' chips and fragmented organic remains. Some of the forms are suggestive of internal molds of small brachiopod shell fragments. Others closely resemble accretionary lapilli (Moore and Peck, 1962), but their epidote-plagioclase assemblage represents an unlikely volcanic composition.

Secondly, the chemical composition of metapelite layers throughout the formation differs noticeably from "average Paleozoic shale" (Pettijohn, 1949) and from many analyses of pelitic slate, schist, and gneiss (note comments at beginning of preceding section on metamorphism). Analyses of several pelitic layers in the formation from chlorite grade to K-feldspar-cordierite grade indicate that combined total iron (as FeO) and MgO are higher than the average shale, CaO is somewhat higher, even though carbonates are lacking, and SiO_2 is lower than average. Mafic and other volcanic rocks are absent in the formation, but the differences mentioned above with regard to Dead River metapelite lie in the direction that would be expected if mafic volcanic detritus were mixed with 'average shale' (Table 1). Table 1 shows, for comparison, the proportions of oxides from fresh basaltic glass to be mixed with those of average shale to approach the percentages in the Dead River metapelite. The mixing-proportions for the basaltic oxides are not equivalent, nor should we expect them to be. Widespread availability of fresh basaltic detritus would have required a regional distribution of aquagene tuff throughout the time of deposition of the formation. This requirement is unrealistic. The differences in the mixing proportions in Table 1 imply a chlorite - iron oxide-rich source. The well-altered products of mafic andesite and basalt, plus small proportions of silt could have constituted this source. Rather uniform mixing by resedimentation could have been achieved by basinward dispersal of suspensions involving the two types of source material. Provenances of maturely weathered, continental debris, and basaltic crust or volcanic islands are considered. One stage, or cycle, of transportation and deposition from one or more sources seems incapable of producing the uniformity of magnesian pelitic composition on a regional scale. The single stage hypothesis is also inadequate in accounting for the compositional range of the coarser clastic fractions that gave rise to quartzwacke and graywacke associated with the pelite in the resulting flysch.

The increasing activity of slope and basin tectonics reflected in the primary sedimentary structures of the upper part of the Dead River Formation, represents a penultimate stage of sedimentation thought to be antecedent to the Taconian orogeny.

References

Bouma, A.H., 1962, Sedimentology of some flysch deposits; Elsevier, Amsterdam; 168 p.

Chinner, G.A., and Schairer, J.F., 1962, The grossular-pyrope join in the system $CaO-MgO-Al_2O_3-SiO_2$: Am. Jour. Sci., v. 260, p. 611-634.

Dott, R.H., Jr., and Howard, J.K., 1962, Convolute lamination in non-graded sequences: Jour. Geology, v. 70, p. 114-121.

Dzulynski, Stanislaw, and Walton, E.K., 1965, Sedimentary features of flysch and greywackes; Elsevier, Amsterdam; 274 p.

Gribble, C.D., and O'Hara, M.J., 1967, Interaction of basic magma with pelitic materials: Nature, v. 214, p. 1198-1201.

Kuenen, Ph.H., 1957, Sole-markings of graded graywacke beds: Jour. Geology, v. 65, p. 231-258.

McKee, E.D., 1965, Experiments in ripple lamination: in Middleton, G.V., (ed.), Primary sedimentary structures and their hydrodynamic interpretation: Soc. Econ. Paleontologists and Mineralogists, Spec. Publication 12, p. 66-83.

Moore, J.G., and Peck, D.L., 1962, Accretionary lapilli in volcanic rocks of the western continental United States: Jour. Geology, v. 70, p. 182-193.

Muffler, L.J.P., Short, J.M., Keith, T.E.C., and Smith, V.C., 1969, Chemistry of fresh and altered basaltic glass from the Upper Triassic Hound Island Volcanics, southeastern Alaska: Am. Jour. Sci., v. 267, p. 196-209.

Pettijohn, F.J., 1949, Sedimentary rocks (Second ed.); Harper and Bros., New York; 718 p.

Shelley, David, 1968, Ptygma-like veins in graywacke, mudstone, and low-grade schist from New Zealand: Jour. Geology, v. 76, p. 692-701.

Smith, A.G., 1968, The origin and deformation of some "molar-tooth" structures in the Precambrian Belt-Purcell Supergroup: Jour. Geology, v. 76, p. 426-443.

Winkler, H.G.F., 1967, Petrogenesis of metamorphic rocks: Springer-Verlag, New York, 237 p.

Trip B-1
Road Log

Trip will leave from the Rangeley Inn at 8:00 a.m., and proceed east on Rt. 4 to Phillips; at Phillips, on Rt. 142 to Kingfield. We then follow Rt. 16 east from Kingfield to North New Portland, then north into the Little Bigelow Mountain quadrangle, following the unmarked, paved road to Long Falls. The total distance is 72 miles, allowing ample time for reading the text and loading geological guns. Weather and time permitting, a brief stop will be made at the height of land where the road crosses the northwest margin of the Lexington batholith. Several major geologic features can be perceived to advantage from here.

We will then proceed directly to Long Falls. The one formal "stop" is the 700 feet of continuous exposure across the structure in the gorge of Long Falls. The preceding text is devoted mainly to this section and the great variety of features it exposes. Therefore the descriptive material of the text serves as the log of our passage southeastward up the course of Long Falls.

The trip will conclude allowing for sufficient time to return to Rangeley for the annual dinner.

Metamorphic Petrology, Mineralogy and Polymetamorphism in a Portion of N. W. Maine

by

Charles V. Guidotti
Department of Geology and Geophysics
University of Wisconsin
Madison, Wisconsin 53706

Preliminary Comments: As this trip emphasizes polymetamorphism the subject matter requires development before field stops can be meaningful. Hence, comments on the fabrics and textures and a summary of pertinent mineralogic data is presented.

Acknowledgements: The writer acknowledges the help and encouragement of J. C. Green, D. S. Harwood, D. J. Milton, K. A. Pankiwskyj and J. Fink Warner. R. H. Moench is thanked especially for allowing me to study the petrology of the three quadrangles (Phillips, Rangeley, Rumford) that he has been mapping. Moreover, discussions with Moench made the writer aware of the significance of the staurolite overprint patterns. Discussions with N. Carter of Yale University have also been very helpful in this respect.

Professors M. P. Billings and J. B. Thompson, Jr. of Harvard University originally got me interested in N. W. Maine and guided my early work. Financial support from Harvard at that time is acknowledged.

Recent financial support has come from the Maine Geological Survey (under the direction of R. G. Doyle), University of California at Davis, and N.S.F. Grants GA-406 and GA-1496. This support is greatly appreciated.

Introduction: The area considered includes the Oquossoc, Rangeley, Phillips, Dixfield, Rumford, Old Speck Mountain, Buckfield, Bryant Pond, and Bethel quadrangles (See road log, Fig. 1). It lies at the NNE end of the high grade metamorphic area extending S.W. through much of central and southern New England, (Thompson and Norton, 1968).

Emphasis is on meta-pelites in grades ranging from biotite to Ksp + Sill zone. The trip can be divided into four parts coinciding with specific areas (Areas A, B, C, and D on Fig. I). They are treated separately because of variable amounts of data for each area and possible differences in the metamorphic history. The areas will be seen in the A, B, C, D order.

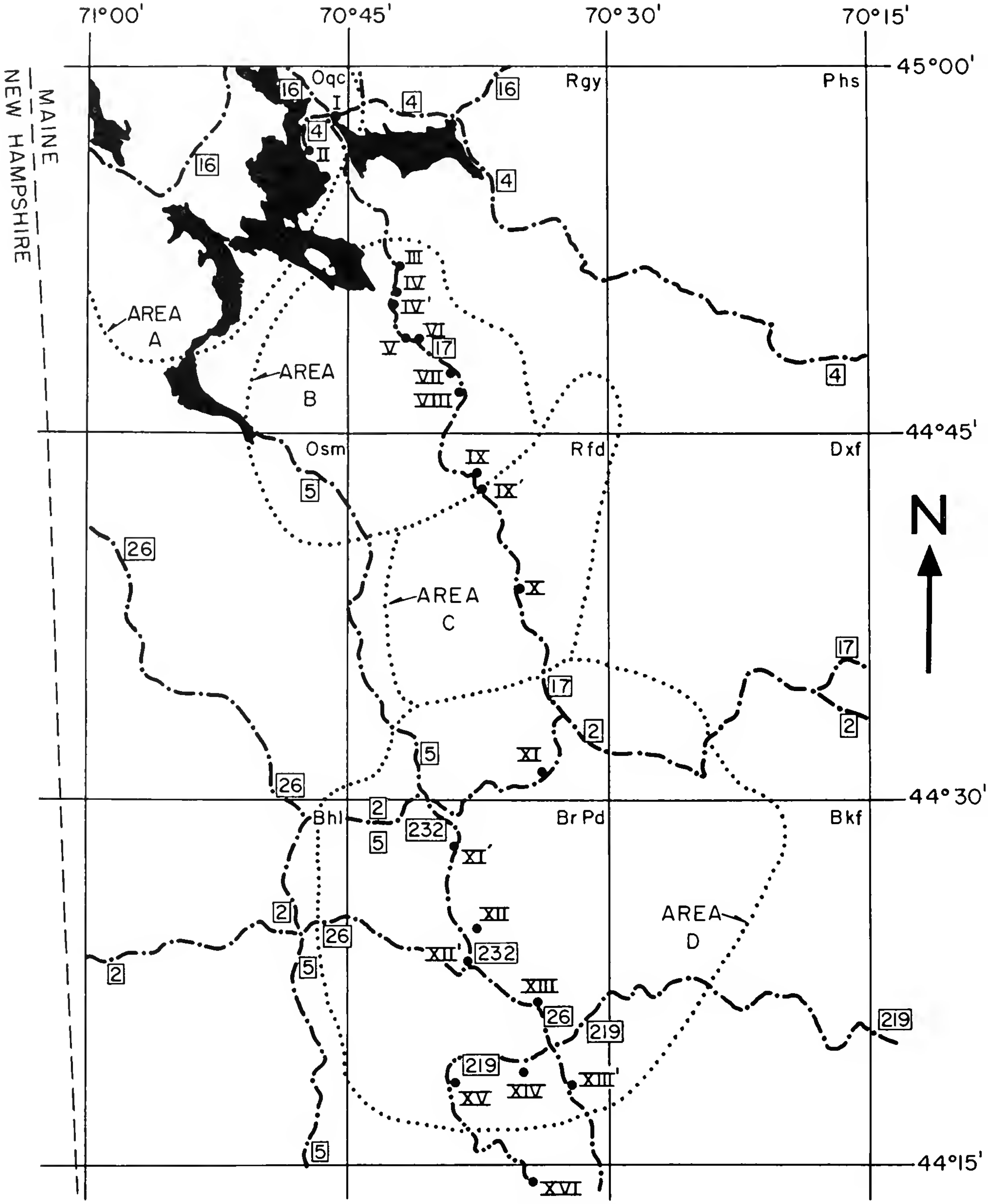
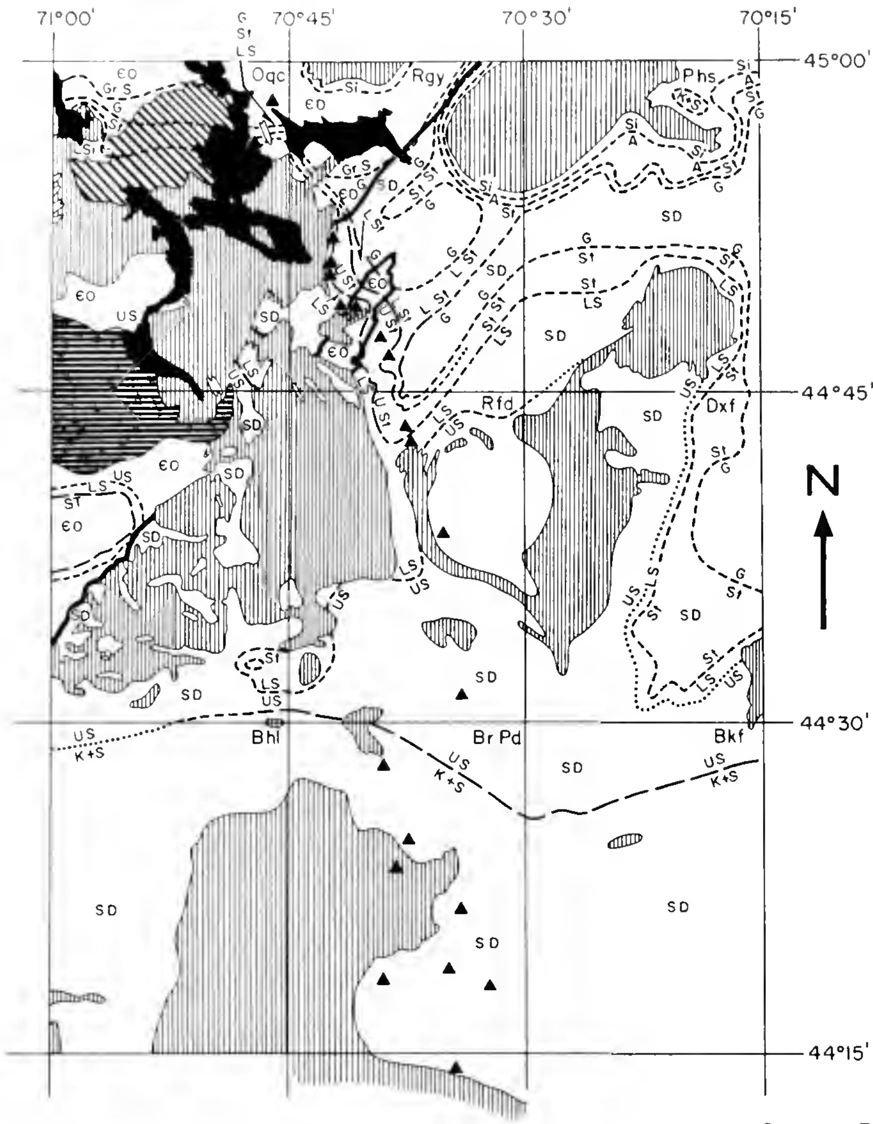


FIG I. ROAD LOG AND LOCATION OF AREAS A,B,C,D.

-ROADS

2 -STATE ROUTE NUMBERS

X -STOPS



0 5
MILES

FIG 2. LOCATION OF ISOGRADS

GRANITIC ROCKS (N.H. MAGMA SERIES)
GRANITE (HIGHLANDCROFT PLUTONIC SERIES)
UMBAGOG GRANODIORITE
▲ ▲ ▲ FIELD TRIP STOPS

ISOGRADS

— — — GOOD CONTROL - - - - - - APPROXIMATE
.... INFERRED
CONTACT CAMBRO-ORDOVICIAN (€O) AND SILURO-DEVONIAN (SD)

METAMORPHIC GRADES

K+S = KSP + SILL
US = UPPER SILLIMANITE ZONE] Si = SILLIMANITE
LS = LOWER SILLIMANITE ZONE]
A = ANDALUSITE St = STAUROLITE
U St = UPPER STAUROLITE
L St = LOWER STAUROLITE
G = GARNET
Gr S = GREEN SCH. FACIES (CHLORITE + BIOTITE)

Objectives include:

(1) Description or tabulation (where possible) of changes in assemblages, modes, and mineral composition in order to define the metamorphic zones.

(2) Observation of typical specimens noting any textural changes with grade.

(3) Use of textures seen in the field (and described from thin section) to document polymetamorphism in Areas B and C.

(4) Showing the distribution and attitude of isograds in much of N. W. Maine.

General Distribution of Isograds: Sketchy data make difficult unequivocal location of the isograds. More problematic is the polymetamorphism mentioned above. Superimposed events may involve disequilibrium, thereby raising questions about the grade present.

Nonetheless, Fig. 2 shows some presumed contemporaneous isograds; the same ones shown in detail for Area B (see Fig. 3). These isograds do not necessarily coincide with those of Moench (this guidebook) because his isograds are drawn to show earlier metamorphic conditions. For example, his staurolite isograd includes rocks with staurolite present as well as those with pseudomorphs indicating the former presence of staurolite.

Sources of data for the isograds in Fig. 2 include:

Oquossoc - Green and Guidotti (1968)

Rangeley - Moench (1966, Pers. Com.); Osberg et al (1968); the writer's work.

Phillips - Osberg et al (1968): the writer's work.

Dixfield - Pankiwskyj (1964); the writer's work.

Rumford - the writer's work.

Old Speck Mountain - Milton (1961) with minor modification by the writer.

Buckfield - Warner (1967) modified by the writer.

Bryant Pond - Evans and Guidotti (1966).

Bethel - Fisher (1962); the writer's work.

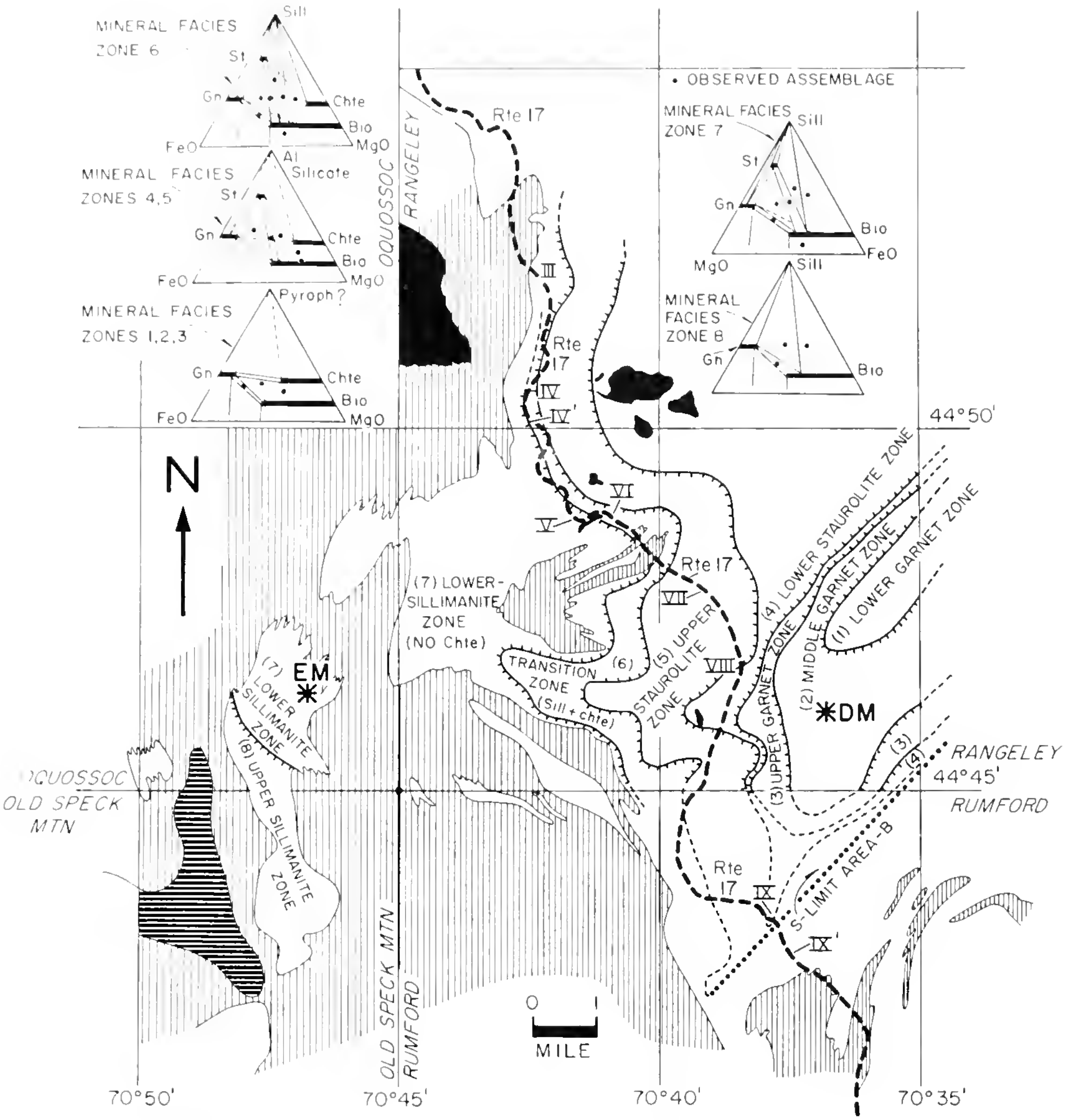


FIG 3. DETAILED ISOGRADS OF AREA B ✳ DM = DOLLY MTN.
✳ EM = ELEPHANT MTN.

Opaque minerals in the pelites of Area A differ from those in B, C, D. Non-graphitic rocks have ilm + Py + Mag ± Pyrr and graphitic ones ilm + sulfides + graphite. Magnetite has not been found in rocks from Areas B, C, D. Thus, at a given P, T, it is likely that PO_2 (and PH_2O) in Area A differed from that in the other areas. Possibly reflecting a higher PO_2 in Area A, biotite has colors in shades of greenish brown, contrasting with the orange to reddish brown (depending upon grade) in rocks from Areas B, C, and D.

AREA - B: Extensive work has been carried out in this area by the writer (including some in progress with A. L. Albee) and by Moench (1966 and pers. comm.) Most textural and fabric data relative to areas east and northeast of Area B are based upon the work of Moench. The isograds shown on Fig. 3 were developed by the last metamorphic event to affect Area B. We will try to show the following about these isograds.

(1) The grade ranges from garnet zone (Dolly Mtn) to upper sillimanite zone (west slope of Elephant Mountain).

(2) Equilibrium has been approached.

(3) The last metamorphism has been superimposed on an earlier metamorphic terrane involving staurolite and andalusite-bearing rocks.

Stratified rocks of Siluro-Devonian age (and some Ordovician) include pelitic schists, biotite granulites, conglomerate and some calc-silicates, (Moench, 1966 and 1969; Osberg et al, 1968). As seen on the map in the introduction (this volume) the beds are deformed into N. E. trending folds and intruded by two-mica adamellite in the western parts of the area.

Several S-surfaces occur in these rocks and can be related to events involving deformation and/or metamorphism. Moench (1966) described 3 surfaces readily recognized in Area B.

(1) S_1 = bedding, now deformed by large N. E. trending folds.

(2) S_2 = an older foliation ranging from a slaty cleavage to coarsely foliated schistosity depending upon grade. It reflects strong dimensional and crystallographic orientation of minerals and seems to be related to the formation of the large N. E. trending folds although recrystallization may be later.

(3) S_3 = a younger foliation which is especially well developed along the western part of the Rangeley quadrangle but also occurs throughout much of the remaining parts of Area B. S_3 consists of a slip cleavage (White, 1949) in low grades and a schistosity in higher grades. Near the western border of the Rangeley quadrangle it strikes N. W. and dips gently N. E.

AREA A: Small amounts of data are available for this area. Stratified rocks belong to the ϵO sequence of the Boundary Mountains Anticlinorium. (Green and Guidotti, 1968; Harwood et al, Trip A-3, this conference)

Meta-volcanics include typical greenstones at low grades and amphibolites at high grades. Low grade meta-pelites include green phyllites and carbonaceous, sulfide-bearing, dark grey slates and phyllites. At higher grades andalusite, sillimanite, and staurolite occur. Paragonite, chloritoid, and pyrophyllite are not found at any grade presumably due to only moderate amounts of Al_2O_3. Stratigraphic equivalents of some of these units e.g. Aziscohos fm (N. Maine) $\approx$ Stowe fm in northern Vermont) do contain paragonite and chloritoid at the appropriate grades (Albee, 1968).

Metamorphic grade ranges from chlorite to upper sillimanite zone. Because of inadequate data only approximate isograds for garnet, staurolite and sillimanite are shown. These isograds are extensions of similar isograds rigorously located by Green (1963A, 1964) in the Errol area. Possibly they correspond with similar isograds in Area B.

Pelitic mineral assemblages are typical (for a given grade) of rocks plotting below the garnet-chlorite join of an AKFm projection except that instead of three phase assemblages, four and five phase assemblages are common, thereby suggesting some disequilibrium. Moreover, chlorite in the staurolite and lower sillimanite zone occurs as radiating aggregates (to 1/2 cm) commonly with an interleaving of anormalous brown and blue layers--suggesting some retrograde effects.

A contact aureole developed around the coarse, porphyritic, Ordovician granite (X on Fig. 2) has been partially recrystallized. Textures and mineralogy of the granite also reflect metamorphism. For example. mortar textures, and minerals such as chlorite, epidote (and/or zoisite), and sericitized feldspar are common in the granite in the garnet zone whereas in the sillimanite zone it consists of "granite" mineralogy and porphyritic texture. In all grades the Ksp is maximum microcline.

Mineral assemblages in Area A are quite similar to those in Area B suggesting a similar facies series (Miyashiro, 1961), i.e. low pressure intermediate. Contact aureoles around some Acadian plutons intrusive into low grade rocks (Green, 1963A, Harwood and Larson, 1969, and N. W. Oquossoc) may represent a lower pressure metamorphism but this is not yet established. Cordierite is quite common in such aureoles.

Preliminary basal spacing work on muscovite shows the most sodic muscovite in pelites of staurolite grade just as in rocks of Area B.

In places S_3 is more pronounced than S_2. Commonly the muscovite laths showing S_3 superimposed upon S_2 have been polygonized (Zwart, 1962). Clearly it represents a shearing event post-dating the formation of S_2.

Garnet zone (See Fig. 3) rocks are fine-grained, greenish-grey phyllites (ignoring the Smalls Falls fm which is black due to graphite and disseminated sulfide) containing qtz + plag + bio + musc + chte + garn (See Table 1 for estimated modes etc.).

Textural features include:

(1) Biotite as isolated 1-2mm tablets.

(2) Garnet as 3-4mm euhedra commonly rimmed epitaxially by coarse laths of Fe-chlorite.

(3) Muscovite, chlorite, quartz, and plagioclase make up the groundmass and orientation of these minerals defines S_2.

(4) Aggregates of medium grained muscovite and chlorite form pseudomorphs after staurolite in some parts of the garnet zone. The pseudomorphs as well as garnets mentioned above cross cut S_2.

(5) S_3 is weak to moderately well developed. The Muscovite laths defining S_3 are partially polygonized. S_3 is generally truncated by the pseudomorphs after staurolite and the garnets although sometimes partially bending around these features. Biotite tablets and coarser laths of chlorite tend to line up along S_3 but nonetheless truncate it as shown by graphite trains within plates of biotite or chlorite. Quartz inclusions occasionally follow the graphite trains in the chlorite. In no case does S_3 deform pseudomorphs after staurolite, coarse chlorite rimming garnet, biotite tablets, or the isolated coarser laths of chlorite.

Some of the garnet zone was once in the staurolite zone but delicate sedimentary features (especially cross beds and graded beds) and clastic textures are retained in many specimens. Moreover, staurolite pseudomorphs never exceed 1/2 cm in length and are absent from much of the garnet zone (as recognized by Moench 1966). Probably much of this zone has never been much more highly metamorphosed than at present.

In the lower staurolite zone (Fig. 3) textures become coarser and specimens then develop a white luster. Isolated laths of chlorite and biotite are distinctly coarser than groundmass muscovite. Garnet, subhedral to euhedral is not rimmed by chlorite. Staurolite occurs as (1) subhedral, moderately to highly poikilitic megacrysts, (2) aggregates of oriented subhedra pseudomorphous after larger staurolite crystals, and (3) anhedral masses enclosed by coarse muscovite and chlorite (this mode being most common). Estimated modes are given in Table I.

S_3 is better developed than in the garnet zone, occasionally even supplanting S_2 as the most prominent foliation. In addition to the textural relationships of S_3 to pseudomorphs etc. seen in the garnet zone, fresh staurolite also truncates S_3. Curved patterns of quartz inclusions in the staurolite shows that staurolite overprints the pattern of S_3 in the groundmass. In no case does S_3 deform muscovite and chlorite partially pseudomorphing staurolite.

The upper staurolite zone (Fig. 3) is characterized by somewhat coarser grain size than the lower staurolite zone and a decrease in modal chlorite (See Table I). Most lower staurolite zone textures are repeated in the upper staurolite zone. Staurolite occurs mainly as types (1) and (2) described in the lower staurolite zone.

Also present in some specimens are large, irregular, highly poikilitic to skeletal andalusite crystals, in some cases almost forming a groundmass (similar andalusite is rare in the lower staurolite zone). The pattern of quartz inclusions reflects S_3 just as commonly seen in staurolite. In many cases the andalusite is partially replaced by coarse muscovite.

The transition zone* is marked by (see Fig. 3 and modes in Table 1):

(1) Appearance of sillimanite.

(2) Persistence of only traces of chlorite.

(3) Some replacement of staurolite by coarse muscovite alone.

(4) Some biotite epitaxial rimming of garnet.

(5) Strong development of S_3 - commonly being more prominent than S_2.

Most other textural features seen in the upper staurolite zone persist into the transition zone.

Lower sillimanite zone (Fig. 3 and modes in Table 1) rocks are similar to the transition zone but grain size coarsens and staurolite is more replaced by <u>coarse</u> muscovite. Concommittent relocation of groundmass muscovite into pseudomorphs after staurolite results in a darkening of color of the rock (Guidotti, 1968). Muscovite replacement of andalusite is complete about half way through the lower sillimanite zone. Epitaxial biotite rims on garnet are most pronounced in this grade.

---

* "transition" referring to the appearance of sillimanite and persistence of chlorite in rocks apparently plotting below the staurolite-chlorite line in an AKFm diagram.

In the upper sillimanite zone (Fig. 3 and modes in Table 1) staurolite is absent (Guidotti, 1970). Textural changes worthy of note here are:

(1) Complete replacement of staurolite by muscovite.

(2) eventual coalescence of muscovite into 1 cm megacrysts.

(3) further darkening of color of hand specimens.

(4) biotite occurs as folia consisting of aggregates of plates.

(5) S_3 is almost totally destroyed by recrystallization.

In summary, an especially important point is that in the appropriate grades staurolite and andalusite have over printed S_3. In a few cases S_3 partially bends around staurolite crystals. But even then inclusion patterns inside the staurolite still reflects an over printing. This point will be considered further below.

Excluding andalusite, the mineralogy, modal changes shown in Table 1, and many of the textural features described above can be integrated into a sequence of metamorphic steps ranging from garnet to upper sillimanite zone. The mineralogy of the various grades can readily be related by means of a series of equations such as*:

Garnet Zone:
(1) Fe Chte + Musc $\rightleftharpoons$ Garn + Bio + Mg Chte + Na-richer Musc + H_2O

Garnet   Staurolite Zone:
(2) Mg Chte + Garn + Musc $\rightleftharpoons$ Staur + Bio + Na richer Musc + H_2O + Qtz

Staurolite Zone:
(3) Mg Chte + Musc $\rightleftharpoons$ Staur + Bio + Na-richer musc + H_2O

Staurolite   Lower Sillimanite Zone:
(4) Staur + Mg Chte + Na Musc $\rightleftharpoons$ Sill + Bio + K-richer Musc + Ab + Qtz + H_2O

Lower Sillimanite Zone:
(5) Staur + Sodic Musc + Qtz $\rightleftharpoons$ Sill + Bio + Fe-richer Staur + K-richer Musc + H_2O + Ab

Lower Sillimanite   Upper Sillimanite Zone:
(6) Staur + Sodic Musc + Qtz $\rightleftharpoons$ Sill + Bio + K-richer Musc + Ab + Garn + H_2O

---

* Only reactions for the lower sillimanite zone and higher have been considered in detail so far. Also Zn in staurolite is not considered here.

Upper Sillimanite Zone:

(7) Na Musc + Garn $\rightleftharpoons$ Bio + Sill + Ab + K-richer Musc + H_2O

Fig. 3 includes AKFm diagrams of the mineral facies for the various grades. Clearly from Table 1 and Fig. 3, some of the zones are based upon continuous reactions, (i.e. the same mineral facies for more than one zone).

Features suggesting that the proposed sequence is real and approximates an equilibrium event are:

(1) The systematic modal, textural, and grain-size changes expected for a continuous metamorphic sequence, ranging from garnet to upper sillimanite zone, are clearly present.

(2) Systematic mineral changes can be mapped as isograds and represented by topology changes of AKFm diagrams.

(3) Systematic compositional changes occur for solid solution minerals with change in metamorphic grade. Muscovite and chlorite are especially noteable. Table 1 shows the changes in paragonite content in muscovite (including only data from the most Al-rich specimens as indicated by the assemblage present). The most sodic muscovite occurs in the upper staurolite zone (also observed by Cipriani et al, 1968) which is reasonably consistent with the experimentally determined muscovite-paragonite join of Eugster and Yoder (1955).

Chlorite exhibits changes in optical properties which indicate, (using Criteria of Albee, 1962), Fe-chlorite in the assemblage garn + bio + chlor in the lower and middle garnet zones, (Fig. 3) (this coincides approximately with rocks containing garnets rimmed by chlorite), and Mg - richer in the upper garnet zone. In staurolite grade rocks chlorite is always Mg-rich.

On Fig. 3 the staurolite and garnet zones are thus subdivided on the basis of continuous changes in the composition of muscovite and/or chlorite.

(4) The distribution of the higher grade isograds (Fig. 3) shows a distinct spatial relation to the two-mica adamellite intrusives*.

Only the andalusite does not fit into the metamorphic sequence outlined above. As will be seen below it can be considered as a relic from an earlier metamorphic event.

---

* Moreover, the higher grade isogradic surfaces have low dips, thereby reflecting the low dipping sheets of two-mica adamellite.

It is of utmost importance to note that an intrinsic part of the proposed metamorphic sequence is the development of various types of pseudomorphs. They clearly relate to the proposed metamorphic sequence because the types and degrees of pseudomorphing are keyed to the metamorphic zones which in turn are based upon modal and mineralogical changes. (See Guidotti, 1968 for details of one of these pseudomorphic features). Moreover, platy minerals within the various pseudomorphs have no preferred orientation and in the appropriate grades include and are intergrown with well developed sillimanite. There can be little doubt that metamorphism producing the isograds of Fig. 3 is responsible for the numerous pseudomorphic textures. Certainly the sharpness (i.e. euhedral) of many of the pseudomorphic forms (e.g. Stop # VIII or Fig 5 in Guidotti, 1970) would be inconsistent with the pseudomorphs being relics from any earlier events.

Data for muscovite corroberates the preceding conclusion in lower grade rocks where sillimanite is absent. One might interpret the pseudomorphing there as random, late retrograde products. However, in the staurolite zone, muscovite from rocks with any staurolite present (i.e. only partial pseudomorphs) is invariably more sodic than that from specimens with staurolite totally pseudomorphed. This indicates that the pseudomorphs are not random late retrograde products, but formed during a distinct, broad-scale, coherent event. Although the pseudomorphing in the lower staurolite and garnet grades is indeed a retrograde event, it is nonetheless produced by the same metamorphic event which developed sillimanite grade rocks to the west*.

Hence, the several grades shown in Fig. 3 appear to have approached equilibrium and are thus "geared" to a single metamorphic event ranging from garnet to upper sillimanite zone--a metamorphic event commonly producing pseudomorphic textures. It is then evident that the overprint patterns of S_3 within staurolite and andalusite grains and pseudomorph development on the outer parts of these same minerals implies that most of these minerals are relics from an earlier generation. Chemical data on the staurolite (work in progress with A. L. Albee) shows that the staurolite has nonetheless readjusted compositionally to relate to the isograds of Fig. 3. With regard to the relic andalusite, its lack of any systematic distribution relative to the isograds shown in Fig. 3 suggests that it is a metastable relic. Commonly the aluminum silicates have been supposed to act in a recalcitrant fashion with regard to reactions which should destroy them. Some minor, "new" staurolite (and possibly even some andalusite) may have formed during the metamorphic event outlined above but this is difficult to prove. Possibly the staurolite occurring as euhedral swarms within larger pseudomorphs represents staurolite formed during this metamorphic event.

---

* The systematic changes in the composition of muscovite from grade to grade are, of course, the best indication that the whole region has been subjected to the same final, metamorphic event.

In the light of the above picture, it is possible to suggest the following sequence of metamorphic and tectonic events in Area B.

(1) Folding along NE trending axes accompanied by the formation of an S_2 cleavage (D_1)

(2) Moderate recrystallization at low grades converting S_2 cleavage into an S_2 schistosity. (M_1).

(3) Renewed tectonic activity forming a low dipping S_3 slip cleavage and accompanied by some minor folding, (D_2). Overlapping and outlasting the tectonic activity, much of the area was metamorphosed to an assemblage of andalusite + staurolite + biotite (M_2). S_3 was recrystallized to a good schistosity.

(4) As a distinctly later event, intrusion of adamellite occurred. Intrusions were concentrated in areas where prominent development of S_3 provided easier access for the magma. The attitudes of the intrusives were controlled by the attitudes of S_3. Accompanying + outlasting the intrusives was another strong metamorphic event (M_3) having a distinct spacial relationship with the distribution of adamellites. Near the areas of abundant intrusives, sillimanite grade was reached thereby effecting a prograde metamorphism of the andalusite + staurolite grade rocks. Further from the intrusives the grade was only garnet zone, thereby retrograding the earlier andalusite + staurolite rocks.

Little evidence exists for events post-dating M_3 in Area B. Only rarely are any layer silicates cut by kink bands--(a distinct contrast with Area C). Minor chlorite after biotite in lower and upper sillimanite grade rocks is related to weathering or obvious zones of faulting.

Preceding M_3 was an event M_2 producing most of the andalusite and staurolite which was later partially pseudomorphed by M_3. Staurolite and andalusite of M_2 truncates and overprints S_2 and S_3 patterns although S_3 occasionally tends to bend around megacrysts (or pseudomorphs) of staurolite. This suggests that most of M_2 post-dates D_2 (which produced S_3) but nonetheless some S_3 shearing persisted somewhat into M_2. The chlorite and biotite plates truncating S_3 could have formed in M_2 (after all D_2 ended) but the relation of their chemistry and modal amounts to the isograds of M_3 suggests that they formed during M_3.

It appears that much of the area in Fig. 3 attained andalusite-staurolite grades (due to M_2) except for parts of Dolly Mountain as mentioned earlier. Preceding D_2 it is obvious there must have been an earlier deformation (D_1) which produced S_2.

Moench (1966, P 1450 and 1452, and pers. comm.) have postulated a significant staurolite and andalusite forming event (I have called it M_1) preceding S_3 in the central part of the Rangeley-Phillips area. Moench, on the other hand, has not distinguished <u>two</u> metamorphisms

post-dating S_3. The writer's work in Area B neither refutes nor confirms Moench's suggestion but several facts suggest that most of the staurolite (and andalusite) in Area B post-dates S_3.

(1) Most pseudomorphs and partial pseudomorphs from garnet to lower sillimanite zone have well preserved shapes - usually being sub - to euhedral. The pseudomorphs (as well as non-pseudomorphed staurolite, garnet etc.) cross cut S_3 showing clearly that the pre-curser minerals post date S_3 also. In Area B it is difficult indeed to demonstrate any staurolite preceding S_3.

(2) Even in some lower staurolite zone specimens the pattern of quartz inclusions in staurolite remnants within partial pseudomorphs have weak patterns showing the overprinting of S_3* (these patterns being very well developed in the upper staurolite and higher grades). In a few specimens from the western slopes of Dolly Mountain (just barely above garnet grade) the S_3 overprinting is as well developed as that in any specimen from the more westerly parts of the Rangeley area. As most rocks have staurolite crystals or pseudomorphs of similar size it seems unreasonable to ascribe some staurolite to an M_1 event since most of it is clearly due to M_2.

Thus, prior to D_2, Area B may have been metamorphosed only to the extent of recrystallizing and coarsening S_2 (indicated as M_1 in our suggested sequence of metamorphic and tectonic events, P).

Still not clearly demonstrated is the implied, distinct separation of M_2 and M_3. Several points bearing on this are:

(1) Cordierite is present in several transition zone rocks and shows overprint patterns on S_3. However it is always strongly rimmed by coarse chlorite and muscovite.

(2) The isograds of M_3--especially those separating the upper staurolite, transition and lower sillimanite zone show no relation to the distribution of andalusite. The vast majority of andalusite observed shows overprint patterns on S_3 and is also significantly replaced by coarse muscovite. This holds true for rocks ranging from lower staurolite zone to well up in the lower sillimanite zone.

Absolutely no relationship exists between andalusite and the isograd which first brings in sillimanite. This isograd clearly involves the breaking of the staurolite-chlorite join on an AKFm diagram to form the sillimanite-biotite join and is associated with several systematic modal and mineralogic changes of the solid solution phases. If the reaction were really just a transition (whether stable or metastable is irrelevant) of andalusite to sillimanite, it is difficult to understand

---

* Obviously the weaker development of S_3 in these lower grade rocks means that S_3 overprint patterns will also be less distinct.

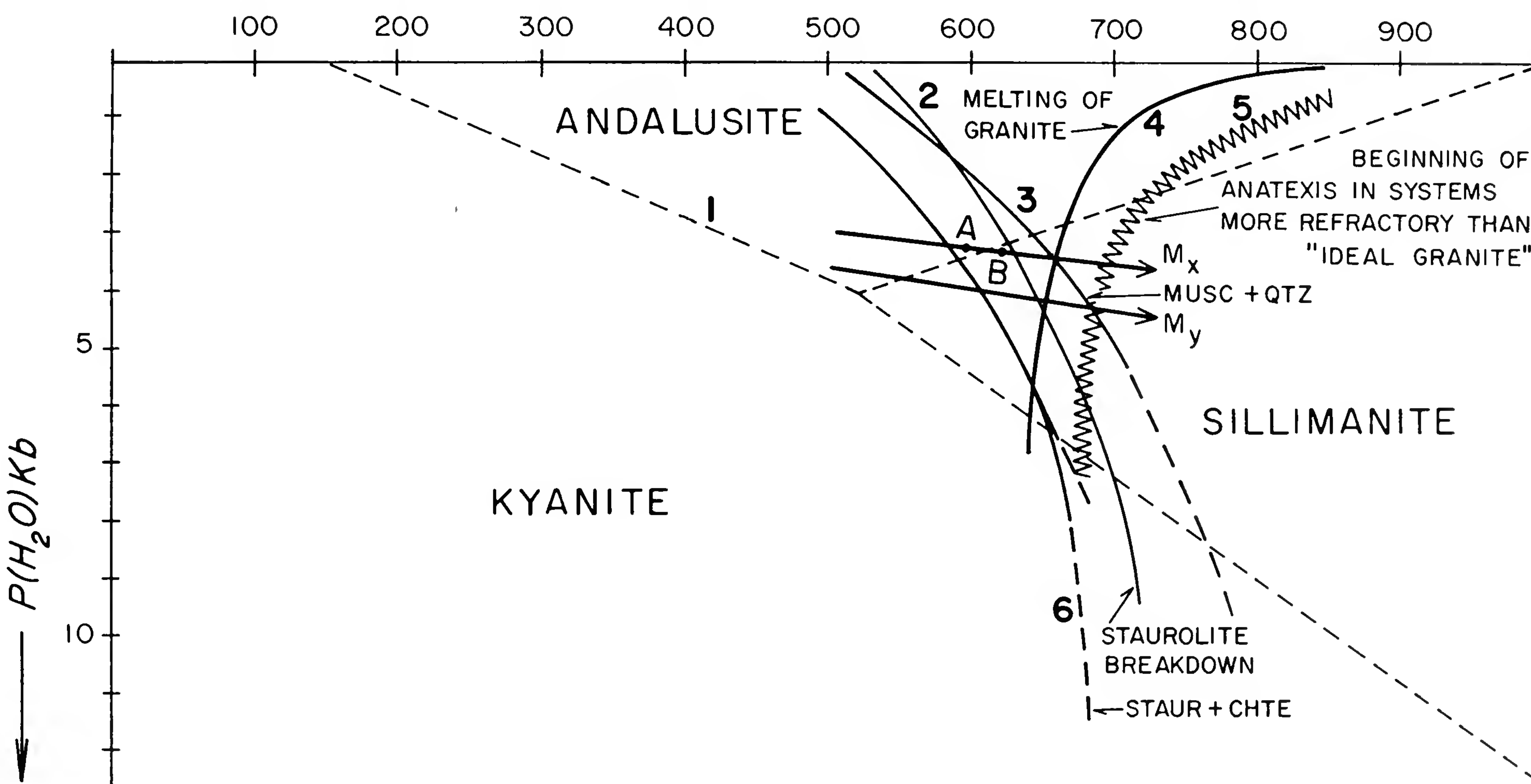


FIGURE 4. PT CURVES GERMANE TO THE TRANSITION FROM STAUROLITE TO UPPER SILLIMANITE ZONE. CURVES INCLUDED ARE: **1** ALUMINUM SILICATES, NEWTON (1966), **2** BREAKDOWN OF STAUROLITE, HOSCHEK (1968), **3** BREAKDOWN OF MUSCOVITE, EVANS (1965), **4** MINIMUM MELTING OF GRANITE, TUTTLE AND BOWEN (1958), **5** BEGINNING OF ANATEXIS, WINKLER (1967), **6** UPPER STABILITY OF THE PAIR, STAUROLITE + CHLORITE

why this should coincide with the other observed modal and mineralogical changes which occur at the isograd. These several considerations are more consistent with the hypothesis that andalusite is merely a metastable relic from M_2. Clearly the production of sillimanite and the various pseudomorphs is spatially related to the plutonic rocks whereas andalusite shows no such relationship.

(3) A hinge effect is present for M_3 such that some rocks affected by M_2 have been retrograded to staurolite and garnet zones whereas others have been prograded to sillimanite zone. (A similar hinge effect was found in the Dixfield areas, (Pankiwskyj, 1964).

Points 1-3 are inconsistent with M_2 and M_3 as part of a single event such as the Mx arrow in Fig. 4. This arrow has an A and B part representing a continuous rise of temperature with time. Instead two separate events (Mx and My arrows) are necessary with My at a somewhat greater pressure (and Mx not surpassing the temperature of point A). Surely the "hinge effect" mentioned above is irreconcilable with M_2 and M_3 as part of one event because rocks in the lower staurolite zone (as an example) would have to be heated up to some temperature T_1 and recrystallized and then cooled to T_2 accompanied by further recrystallization, (all within the same event) during which time temperatures to the west (and part of the same thermal regime as in the lower staurolite zone rocks) continued rising to produce sillimanite. Clearly this is unreasonable. Two separate events M_2 and M_3 are required. Finally, even the isograds of M_3 can not be attributed merely to the late cooling history of the plutons as shown by the fact that the adamellite truncates the isograds in several places (See Fig. 3). Clearly, the isograds of M_3 are contemporaneous with the emplacement of plutons but the relations described indicate that plutons and isograds are due to a common thermal source rather than one controlling the other.

Inasmuch as M_3 and the adamellite intrusions are contemporaneous, the fact that M_3 post dates S_3 suggests that <u>the formation</u> of S_3 is not the result of intrusion of the adamellites*. The distribution of adamellite <u>is</u> related to S_3 (as pointed out by Moench also) because the intrusions commonly occur as thick, tabular bodies parallel to S_3 in the Oquossoc and S. W. Rangeley areas. But the increased intensity of S_3 near plutons noted by Moench (1966) may not result from the plutons; instead the distribution of plutons may be controlled by the degree of development of S_3.

---

* Moreover, it is hard to visualize how an intruding pluton could produce the almost horizontal S_3 surfaces seen along the W. side of the Rangeley area.

Inasmuch as M_3 is related to the adamellites and the adamellites belong to the New Hampshire Plutonic series, (Green, 1964; Green and Guidotti, 1968) which is Acadian in age, then M_3 must be the same age. M_2 must also be at least Siluro-Devonian in age. Hence we have two Siluro-Devonian metamorphisms.

AREA - C: Only moderate amounts of work have been done here so far. The stratified rocks are at least partly equivalent to those in Area B. Textures, fabrics and mineralogy matching S_2, S_3, M_2, and M_3 of Area B can be recognized in the northern sections of Area C. Moreover, the approximately located isograds for staurolite and sillimanite zones are commonly almost horizontal just as in Area B. These observations suggest Areas B and C have been effected by the same events. Even the assemblages developed appear similar to those of M_2 and M_3 in Area B.

However, most striking now, in rocks of Area C is the abundant evidence for disequilibrium.

(1) Many assemblages contain excessive numbers of phases in terms of the phase rule -- e.g. Sill + Bio + Garn + Staur + Chte is common in the N and NE sections. Moreover, the anomalous blue color of chlorite suggests Fe chlorite which is unexpected in high grade pelites.

(2) Biotite is commonly replaced by chlorite and sagenitic rutile although the same rock may also contain clean megacrysts of chlorite larger than any of the biotite plates. In some cases, chlorite of two different colors is present.

(3) Sillimanite is sometimes much replaced by sericitic mica. Most commonly it is resorbed so that it now occurs mainly as inclusions in quartz, especially quartz poikilitically included in megacrysts of muscovite. Concommittent with resorption of sillimanite, muscovite megacrysts commonly fragment into sericitic aggregates. A distinct antipathy exists between sillimanite and biotite; a marked contrast with their stable textural relations at high grades.

(4) K/Na values of muscovite (using basal spacings) show little relation to assemblage present. Moreover, little or no systematic relation between isograds and K/Na values of muscovite can be discerned. The quite systematic relation of K/Na in muscovite and grade which occurred in Area B ends abruptly along an ENE trending line just south of Byron (See Fig. 3).

(5) A widely spaced but distinct slip or shear has affected many rocks appearing as sharp, angular kinks in all layer silicates (including coarse muscovite and chlorite which sometimes form partial pseudomorphs after staurolite). Kinking is most pronounced (and sometimes accompanied by mortar structure in quartz) in the area to the SE and E of Byron. In the vicinity of Thomas Farm Brook a breccia zone occurs.

(6) In the N. Central section of Area C, garnet and staurolite commonly have thin rims of fine grained, anomalous blue chlorite and sericite. The rims also follow along cracks and fractures in the garnet or

staurolite.

Hence, disequilibrium seems certain in much of the area, the evidence suggesting retrograding. Much of the area was upper sillimanite zone but some parts possibly only staurolite grade before the retrograding. The coarse, migmatitic Noisey Brook gneiss (Stop # X) may have been Ksp + Sill zone. At Stop # X, it is a coarse, swirled, "soupy" rock containing pegmatites and small adamellite bodies.

Two additional points possibly helping to relate rocks in Area C to Areas B and D are:

(1) In the N and NE sections commonly occur small, euhedral staurolite crystals (1 mm) or aggregates of crystals growing on grain boundaries of quartz and plagioclase. In some cases the euhedra occur in the same rock with larger, ragged staurolite grains which contain coarse needles of sillimanite. The euhedral staurolite "appears" fresh and as if it formed in a later event.

(2) In the S. part of the area, aggregates of fine grained euhedral sillimanite needles occur on quartz and plagioclase grain boundaries. Such aggregates are present in rocks containing coarser sillimanite which is partially resorbed. Again, it "appears" as if the euhedral needles are the result of a late event.

With these observations and impressions the writer tentatively suggests that the same events affected Parts B and C but with two additional events for the latter. These are: (in order)

(1) A metamorphic event (M_4) retrograded the northern sections to staurolite and possibly garnet zone whereas the southern sections were retrograded only to somewhat lower in the sillimanite zone. Probably equilibrium was not attained and the coarse chlorite may have formed in this event.

(2) Deformation produced the kink bands cutting the layer silicates. Possibly retrograding of biotite to chlorite + sagenitic rutile as well as the thin rims on garnet and staurolite formed in conjunction with the shearing that produced the kinks.

M_4 could be in Area C the lower grade portion of the Ksp + Sill metamorphism affecting much of the area covered in Area D.

AREA - D: This area includes much of the southern 1/3 of the Rumford quadrangle because the mineralogy of rocks there can be systematically related to the Ksp + Sill isograd. Rocks, presumably Siluro-Devonian in age, are migmatitic throughout the area. Migmatites contain Ksp only on the upgrade side of the Ksp + Sill isograd but the isograd itself is based on the association Ksp + Sill in the groundmass.

Detailed study by Evans and Guidotti (1966) suggested an approach to equilibrium and the following isogradic reaction.

(8) Musc + Qtz + Ab (from plag) $\rightleftharpoons$ K, Na spar + Sill + H_2O

The migmatites presumably formed by partial melting such as suggested by Lundgren (1966) for similar rocks in Connecticut, but no detailed work on this question has been done here. Chemical work by Fisher (1962) in the Bethel area argues against a metasomatic origin for the migmatites. Fisher favored a partial melting origin but also considered metamorphic segregation.

The possibility of getting the four phase assemblage, Sill + Ksp + Plag + Musc were clearly recognized by Thompson (1961) and (Pers. Com. 1960) due to addition of CaO to the AKNa system. Evans and Guidotti (1966) discussed this assemblage in Area D and noted that the muscovite maintained a constant composition (Pg_6) over a broad area south of the isograd. They also found that muscovite from the associated specimens containing only Musc + Sill + Plag was more sodic than muscovite from specimens with the four phase assemblage. Moreover, plagioclase in non-Ksp rocks was commonly more sodic than that in rocks with the four phase assemblage. These observations are not expected in terms of theory.* To explain these observations Evans and Guidotti suggested:

(1) Thermal gradients flattened south of the isograd such that all exposed rocks formed at the same PT.

(2) The fugacity of water was variable from specimen to specimen but internally controlled. For the four phase assemblage it was suggested that fH_2O was buffered by the assemblage.

Evans and Guidotti (1966) favored # 2 as a working model.

Although plagioclase in the four phase assemblage becomes more calcic with increased grade (see Fig. 5 of Evans and Guidotti 1966) the weak zoning present is normal instead of reverse as expected. A possible explanation is that the zoning is a late phenomenon. It may represent late recrystallization in the presence of a silicate melt (i.e. the migmatitic bands). Sodic rims would then result from simple fractional crystallization. In studies by Barker (1962) and Binns (1964) plagioclase in Ksp + Sill rocks does have reverse zoning. Possibly rates of reaction (and cooling) control the nature of zoning in plagioclase from such rocks.

Preliminary work on the Ksp in these rocks (with H. H. Herd of U. C. Davis) using Wright and Stewart (1968) to characterizing the structure state has shown:

(1) Groundmass Ksp coexisting with sillimanite is invariably close to

*Theoretically one expects muscovite in the four phase assemblage to become K enriched with increased grade and the coexisting plagioclase to become enriched in An.

orthoclase.

(2) Coarse grained Ksp from megacrysts and migmatitic bands is commonly more ordered -- averaging near Spencer B but ranging to Spencer U.

(3) Ksp from low alumina rocks such as calc-silicates or biotite granulites is always more ordered -- ranging from Spencer B to maximum microcline.

(4) Ksp from obviously retrograded rocks (e.g. biotite going to blue chlorite and sagenitic rutile) is usually more ordered -- ranging to maximum microcline -- regardless of what the original rock may have been.

An Acadian age is assumed for the metamorphism of Area D because many of the intrusive plutons are similar to the Concord granite, Fisher (1962), Billings (1956). The large pluton south of Area D is also shown by Doyle (1967), Page (1968), Osberg et al (1968) etc. as belonging to the New Hampshire Plutonic Series (Acadian age). Moreover, Page also ascribes the pegmatites intruding the metamorphics to the New Hampshire Plutonic Series. Hence evidence exists for a heating event at this time.

Contradictory data for an Acadian age are two Permian ages (by Rb-Sr) from micas in pegmatites (Paris Hill, S. W. corner of Buckfield area) listed in Faul et al (1963). Suggestion of a Permian metamorphism, if real, raises the possibility that M_4 in Area C represents the northern limit of such an event superimposed on M_3 which has an Acadian age. Such speculation would agree with the "Probable NE limit of Permian metamorphism" in Fig. 2 of Faul et al (1963).

Available age dating in N. W. Maine is too scanty to establish any certainty of ages of metamorphism. Presently the writer assumes an Acadian age for the Ksp + Sill isograd but allows that it could represent an event somewhat younger than M_3 in Area B.

Fig. 4 is a summary diagram indicating supposed conditions of metamorphism for M_2, M_3 and the Ksp + Sill isograd.

References

Albee, A. L. (1962) Relationships between the mineral association, chemical composition and physical properties of the chlorite series: Am. Mineral., 47, 851-870.

__________, (1968) Metamorphic zones in Northern Vermont: in Zen et al, eds. Studies of Appalachian Geology, John Wiley and Sons, New York

Barker, F. (1962) Cordierite-garnet gneiss and associated microcline-rich pegmatite at Sturbridge, Massachusetts, and Union, Connecticut: Am. Mineral., 47, 907-918.

Billings, M. P. (1956) The geology of New Hampshire, Part II - bedrock geology. New Hampshire State Planning and Development Commission, Concord, New Hampshire

Binns, R. A. (1964) Zones of progressive regional metamorphism in the Willyama complex, Broken Hill district, New South Wales. J. Geol. Soc. Australia, 11 pt. 2, 283-330

Cipriani, C., F. P. Sassi, C. V. Bassani (1968) La composizione delle miche chiare in rapporto con le constanti reticolari e col grado metamorfico: Rend. Soc. Miner. Ital., 24, 3-37.

Doyle, R. G. (1967), Preliminary geologic map of Maine: Maine Geol. Surv., Dept. of Econ. Devel., Augusta, Maine.

Eugster, H. P. and H. S. Yoder (1955) The join muscovite -- paragonite: Carnegie Inst. Wash. Year Book, 54, 124-126.

Evans, B. W. and C. V. Guidotti (1966) The sillimanite -- potash feldspar isograd in western Maine, USA: Beitr. Mineral. Petrog., 12, 25-62.

Faul, H., T. W. Stern, H. H. Thomas, and P. L. D. Elmore (1963) Ages of intrusion and metamorphism in the northern Appalachians: Am. Jour. Sci., 261, 1-19.

Fisher, I. S. (1962) Petrology and structure of the Bethel area, Maine: Bull. Geol. Soc. Amer., 73, 1395-1420.

Green, J. C. (1963A) High-level metamorphism of pelitic rocks in northern New Hampshire: Am. Mineral. 48, 991-1023.

__________, (1963B) Alkali metasomatism in a thermal gradient, two possible examples: Jour. Geol., 71, 654-657.

Green, J. C. (1964) Stratigraphy and structure of the Boundary Mountains anticlinorium in the Errol quadrangle, N.H.--Maine: Geol. Soc. Amer., Special Paper 77, 78 p.

__________ and C. V. Guidotti (1968) The Boundary Mountains Anticlinorium in northern New Hampshire and northern Maine: in Zen et al, eds. Studies of Appalachian Geology. John Wiley and Sons, New York.

Guidotti, C. V. (1966) Variations of the basal spacings of muscovite in sillimanite-bearing pelitic schists of northwestern Maine: Am. Mineral., 51, 1778-1786.

__________ (1968) Prograde muscovite pseudomorphs after staurolite in the Rangeley-Oquossoc areas, Maine: Am. Mineral., 53, 1368-1376.

__________ (1970 in press) The mineralogy and petrology of the transition from the lower to upper sillimanite zone in the Oquossoc area, Maine.

Harwood, D. S. and R. R. Carson (1969) Variations in the delta-index of cordierite around the Cupsuptic pluton, west-central Maine: Am. Mineral. 54, 896-908.

Lundgren, L. W. (1966) Muscovite reactions and partial melting in southeastern Connecticut: Jour. Petrol., 7, 421-453.

Milton, D. J. (1961) Geology of the Old Speck Mountain quadrangle, Maine: Ph.D. thesis, Harvard University.

Miyashiro, A. (1961) Evolution of metamorphic belts: Jour. Petrol., 2, No. 3, 277-311

Moench, R. H. (1966) Relation of S_2 schistosity to metamorphosed clastic dikes, Rangeley-Phillips area, Maine: Bull. Geol. Soc. Amer., 77, 1449-1462.

__________ (1969) The Quimby and Greenvale Cove Formations in Western Maine: U. S. Geol. Surv. Bull. 1274-L, P. 1-17.

Osberg, P. H., R. H. Moench, and J. L. Warner (1968) Stratigraphy of the Merrimack Synclinorium in West-Central Maine in Zen et. al. eds. Studies in Appalachian Geology. John Wiley and Sons, New York

Page, L. R. (1968) Devonian Plutonic Rocks in New England: in Zen et. al. eds. Studies in Appalachian Geology. John Wiley and Sons, New York

Pankiwskyj, K. A. (1964) The geology of the Dixfield quadrangle, Maine: Ph.D. Thesis, Harvard University.

Thompson, J. B., Jr. (1961) Mineral facies in pelitic schists: in Russian, English summary; in Physico-chemical Problems of the Formation of Rocks and Ores, G. A. Sokolov, ed., Akad, Nauk SSSR.

__________________ and S. A. Norton (1968) Paleozoic Regional Metamorphism in New England and Adjacent Areas: in Zen et. al. eds. Studies in Appalachian Geology. John Wiley and Sons, New York

Warner, J. L. (1967) Geology of the Buckfield quadrangle, Maine: Ph.D., thesis, Harvard University, 232 P.

White, W. S. (1949) Cleavage in east-central Vermont: Trans. Am. Geophys. Union, 30, No. 4. 587-594

Wright, T. L. and D. B. Stewart (1968) X-ray and optical study of alkali feldspar I. Determination of composition and structural state from refined unit-cell parameters and 2V: Am. Mineral. 54, 38-87.

Zwart, H. J. (1962) On the determination of polymetamorphic mineral associations, and its application to the Bosost area (central Pyrenees): Geol. Rund., 52, 38-65.

Road Log for Trip B-2

Assemble in front of the Town and Lake Motel on Rte. 4, west side of Rangeley. We leave at 8:00 a.m. Head due west on Rte 4.

Mileage

6.7 Stop I in the N.E. part of the Oquossoc quad on Rte. 4 just W. of the junction with Rte. 16. Park on the wide shoulders. The rock is a typical greenstone (nearby pelites indicate biotite zone) containing actin-chlte-epid-albite and some calcite, quartz, sulfides, etc. The protolith is presumed to be basaltic flows but some pyroclasitics also are present in other outcrops. Notice the epidote-rich pods and clots in the otherwise massive to foliated volcanics. Possibly these represent volcanic bombs.

In some cases, remnant plagioclase and pyroxene grains have been found in these rocks.

Drive through Oquossoc village.

8.1 "T" in road--at Haines landing on Mooselookmeguntic. Turn left (South)

10.2 Stop II Tight parking--take care on pulling off to left! Some mud! Main stop is at Adam's Camps and just south where the best crops are. The lead cars will park 50-75 yards S. of Adam's Camps and the others as space is available.

Here we see 3 rock types--all in lower sillimanite zone or upper staurolite zone.

(1) Quartz pod pelitic schist of the Deer Mountain member of the Albee fm.

(2) Meta volcanics (amphibolites here) commonly interbedded with the pelites.

(3) Metamorphosed granite of the Highlandcroft Plutonic Series. Sometimes has Pink Ksp.

In the pelites one can find good 1/4" euhedral garnets and tourmaline; occasionally staurolite is also seen in hand specimen. Sillimanite is always microscopic. Notice some of the coarse clots of chlorite. (see text)

Continue S. on dirt road.

12.2 Turn left--still on a dirt road.

13.4 Dirt road meets Rte. 17 (Paved). Turn right (south).

18.4 Stop III--on Rte 17. Fair amount of parking on both sides of road, but use caution as there are many curves on this road. Most of this crop is cgl and grit but the more aluminous beds have 1/16 to 1/8" staurolite which is partially rimmed by muscovite. Sillimanite is present in thin section. The grade is lower sillimanite zone.

20.2 Stop IV Drill site--parking to the right, left, and just south. The Union Water Power Co. drilled and cored down to about 900 ft. at this site. They allowed me to take a sample every 20 feet or so. At the surface the grade is Transition Zone (i.e. sill + chte). With increased depth the grade rises and at about N 850 feet the hole hits adamellite (which field mapping by Moench could have predicted). About 150 feet from the granite there is an abrupt rise in muscovite and disappearance of sillimanite, suggesting K-metasomatism just as described by Green (1963) near the contacts of similar adamellites in the Errol quad.

20.8 Walk .6 mile S. on Rte 17 to Heights of Land. Crops are just about on the isograd for the lower sillimanite zone. Very good staurolite schists. In a few places, irregular, remnant andalusite weathers to form patches of higher relief.

Note Elephant Mtn. to W.S.W.--Upper sillimanite rocks lie on the west slopes.

22.8 Stop V--Optional--if only a few cars present! Very difficult parking, narrow shoulders. Rocks are merely an example of the dull mineralogy that occurs in low alumina rocks--Only biotite granulites and grits are present although the crops are in the lower sillimanite zone.

23.4 Middle of Beaver Pond.

23.8 Stop VI Just over the crest of the hill east of Beaver Pond. Park on the right side of road-wide shoulders. Walk back to see the interbedded biotite granulites and calc silicates. Rocks at this stop are in the Transition Zone. .2 miles down hill are good dense, white schists. Sillimanite is present in thin section. Some staurolite is totally pseudomorphed by coarse muscovite but some is only rimmed by the muscovite (N. most crop).

25.2 Stop VII Park on the left side of highway--wide shoulders. Rocks are in the upper staurolite zone. Notice the pseudomorphs which now consist of aggregates of euhedra in a matrix of muscovite and quartz.

26.9 Stop VIII Just across the bridge over the Swift River. Plenty of parking but watch embankments on entering parking area. Rocks are low to upper staurolite zone. Follow trail down stream on S. side of River. About 1/8 to 1/4 mile down stream are large slabby crops with very good pseudomorphs (similar to VII) which clearly indicate two periods of staurolite formation.

32.3 Stop IX--at Byron picnic area--in the Rumford quadrangle. Rocks are staurolite grade. Very good development of staurolite! Also notice on some slab surfaces the "turkey track" pseudomorphs after andalusite--again suggesting polymetamorphism. In some specimens andalusite is still present but rimmed by muscovite. It is interesting that here and elsewhere in the Rangeley area andalusite seems best preserved in rocks rich in sulfides and graphite. One can speculate that the fluid in such rocks had a lower % of H_2O and thus reactions removing metastable andalusite (especially by forming hydrous minerals like muscovite) may have been inhibited. Some workers have appealed to dissolved Fe_2O_3 in the andalusite acting as a stabilizer.

The other common site for relic andalusite is in quartz veins. This may result from relative inaccessibility to solutions bearing K_2O.

LUNCH STOP

33.2 Stop IX[1] Optional for later trips using this guide book. To the east in the woods about 1/4 mile can be found good crops of the very sulfitic Smalls Falls formation. Biotite in pelitic specimens is nearly colorless, suggesting phlogopite. It is the only unit that seems to have cordierite fairly commonly.

38.8 Stop X Roxbury Picnic area--Noisey Brook gneiss. Rocks here are superb examples of swirled up "soup". Bedding seems to have been obliterated. Rocks were originally upper sillimanite zone but have now been significantly retrograded. Sillimanite shows many signs of resorption. Crops a little S. of here show some signs of regeneration of sillimanite.

Continue S. on Rte. 17 to the center of Mexico.

Junction of Rte. 17 and Rte. 2 at stop sign in Mexico. Turn right on Rte. 2 and continue across bridge over Swift River.

46.1 Rte. 2 bears left by the stop sign at the State Armory. Follow Route 2 through Rumford.

47.7 --at the pull out beside the Androscoggin River, just west of town. Re-group here. Then proceed W. on Rte. 2.

50.7 Stop XI --big crops in back of Mammoth Mart. Upper sillimanite zone, coarse schists. Notice the clots or eyes of muscovite which are probably pseudomorphous after staurolite. In rocks of higher grade, the muscovite in the pseudomorphs re-crystallizes to single large megacrysts much like occurred at Stop X.

56.5 Junction with Rte 232. Turn S. on Rte 232 across the Androscoggin River.

59.5 Stop XI -- Optional stop for people using this field guide at a later time. Rocks on Barker's High Ledge are Ksp + Sill grade. Here one finds some of the best development of single, subhedral Ksp megacrysts up to 2 inches in good sillimanite-bearing schists.

64.2 Stop XII at N. Woodstock. Turn left on small paved road and park near the end of the paved section (about 1/4 mile). Walk 250 yards up the dirt road beside Billings Hill brook. Then go into the stream bed to see the large crops of migmatites and biotite schists. Most crops of this unit (Billings Hill fm) are fairly rich in sulfides. Try to find some of the more aluminous beds. Some of them are as much as 30% sillimanite which is intergrown with fairly coarse, translucent Ksp. Be careful here, the rocks are very slippery! -- Back to Rte 232 and turn left, to the south and resume mileage clocking.

65.9 -- Intersection with Rte 26 -- turn left and then stop in the big rest area.

Stop XII[1] -- Walk back to granodiorite crop on Rte 232. Different from most of the big Sebago pluton in that it is not per-aluminous. In places it has small amounts of hornblende. Some jointing surfaces have thin seams of epidote. The feldspar to the sides of the seams has commonly become pink in color.

Head S. on Rte 26.

66.3 more of the granodiorite

69.7 Stop XIII at S. Woodstock -- Big crops of Sill + Ksp rocks. Sillimanite is very abundant and megacrysts of Ksp are common. Some of the megacrysts contain subhedral to euhedral, bipyramidal quartz.

If many cars present, some can park .3 miles down the road.

73.9 Stop XIII[1] Auxilliary Stop -- at Snow's Falls picnic area. Good Ksp + Sill migmatites.

Return to Trapp Corner (i.e. N. on Rte 26)

76.1 -- Trapp Corner, turn left (W) on Rte 219 and go through West Paris.

77.5 -- On west side of W. Paris across RR tracks and bridge, Rte 219 bears right -- do not follow it. Head up the hill.

77.8 Stop XIV -- Take a sharp right into the Bell Mineral Quarry and park. Only about 6 cars can fit in here so the others will have to park to the S. on the paved Road. Walk for .6 miles on dirt road to the quarry. THIS IS AN ACTIVE FELDSPAR QUARRY. THE MANAGEMENT ALLOWS PEOPLE IN HERE AT THEIR OWN RISK. PLEASE NOTE: YOU ENTER THIS QUARRY PROPERTY AT YOUR RISK AND THE BELL MINERALS CO. CAN NOT BE HELD LIABLE. PLEASE STAY OUT OF THE "CAVERNS".

Here we can see some spectacular inclusions of calc-silicates "floating" in pegmatite. The diopside has about 30-40% of the hedenbergite end member. Quartz + calcite are common in many specimens. No wollastomite has been found. On some of the loose blocks at the mouth of the quarry look for evidence of migration of components.

Head back toward West Paris.

78.1 Junction with Rte 219 -- Turn left on Rte 219.

82.5 Greenwood Village -- end of Rte 219 -- Turn left by the N. end of Hicks Pond. Head S. along the ponds.

83.3 Stop XV on W. shore of Hicks Pond. Nicely exposed recumbent fold in migmatites. Provides some of the evidence of strongly overturned folding in this region.

Continue S. past Mud Pond etc.

87.1 at Nobles Corner turn left and follow winding dirt and paved road.

90.5 Turn right (South) at the 600' Corner

91.8 Stop XVI -- Park on the right -- Follow small dirt road on left side of road into Crockett Ridge quarry. We are now in the Norway quadrangle. The calc-silicates here show rather strong evidence of element exchange between carbonate beds and biotite granulite with good bands of diopside developing at the interfaces. Some of the bands are 1/4 to 1/2 inch thick.

Table I Mineral Data* for Area B

Assemblages are those with Max #'s of Phases. Metamorphic Zones keyed to Fig 3.	Muscovite			Chlorite(4)			Biotite				Staurolite		Sill.	Plag.	Garn.
	d(002)Å	%Pg.	Modal %	Color (4) X'd Nicols	Mg/Fe	Modal %	B	Color	Mg(3)/Fe	Modal %	Mg/Fe	Modal %	Modal %	An%	Modal %
Zn.1: Gn+Bio+Chte	9.977	9	29	blue	Fe-rich	12	N.D.(5)	orange brown	N.D.	5	X	X	X	ND	Tr
Zn.2: Gn+Bio+Chte	9.970	12	27	blue	Fe-rich	12	N.D.	orange brown	N.D.	8	X	X	X	ND	1
Zn.3: Gn+Bio+Chte	9.954	17	30	blue + brown	Fe+Mg	7	N.D.	orange brown	N.D.	10	X	X	X	ND	1
Zn.4: Staur+Bio+Chte±Gn	9.945	20	27	brown	Mg-rich	5	N.D.	orange brown	N.D.	20	N.D.	3	X	ND	1.5
Zn.5: Staur+Bio+Chte±Gn	9.933	24	26	greenish brown	.9816	2	1.635	orange brown	.8704	20	.1798	4	X	17.5	2
Zn.6: Sill+Staur+Bio+Gn+Tr.Chte	9.943	21	22	greenish grey	X	.2	1.638	moderately dark orange brown	.9005	22	.1875	6	.4	18	1.5
Zn.7: (Rangeley) Sill+Staur+Bio±Garn	9.951	18	22	X	X	X	1.638	dark orange brown	.8763	24	.1845	4	2	20.5	1
Zn.7: (Oquossoc) Sill+Staur+Bio±Garn	9.951	18	20-25	X	X	X	1.642	dark orange brown	.7040	25	.1518	4	5-8	20	1
Zn.8: Sill+Bio+Garn	9.963	14	10-15	X	X	X	1.648	dark reddish brown	.7311	25-30	X	X	10-15	25	1

*(1) Modes are averaged visual estimates.
(2) Compositions etc. are averaged values and thus of only limited use. Detailed values for specific specimens will be published elsewhere.
(3) Mg/Fe, biotite based only on probe data; zones 5, 6, 7 by A. L. Albee, zones 7 and 8 by B.W. Evans.
(4) Data on chlorite from zones 1-4 are based only on optical parameters. Thus Mg/Fe is not designated specifically.
(5) N.D. = No data

TRIP C

Pre-Silurian rocks in the Boundary Mountains anticlinorium, northwestern Maine[1]

by
Eugene L. Boudette
U.S. Geological Survey
Boston, Massachusetts

Introduction

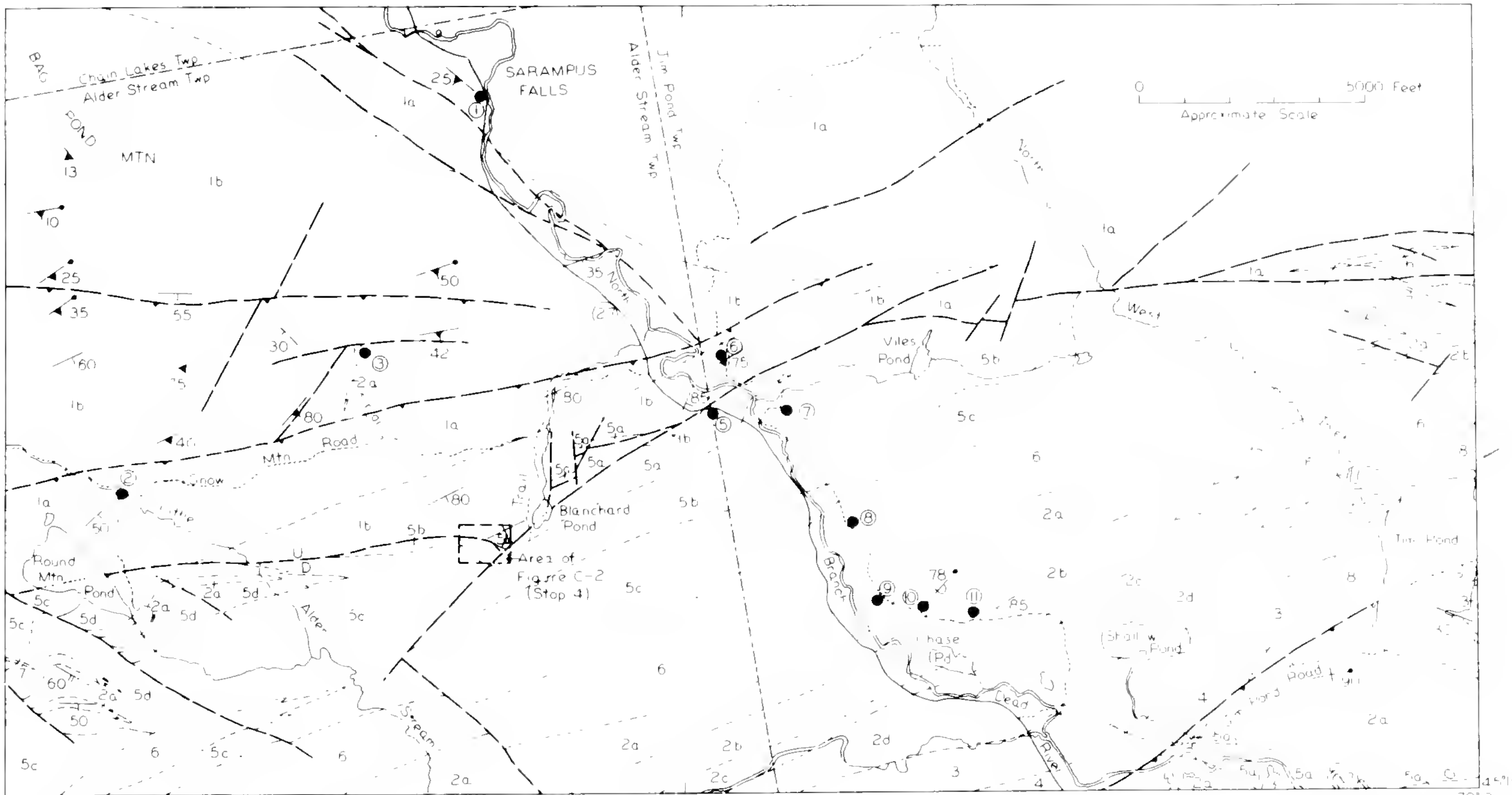
Trip C covers a relatively confined area on the southeastern flank of the Boundary Mountains anticlinorium where Silurian rocks are almost entirely lacking, and where the pre-Silurian rocks constitute an upright, essentially homoclinal succession facing southeast (fig. 1). The purpose of this trip is to examine the evidence that establishes the time and space relationships of an ultramafic complex of the alpine type and granitic rocks that intrude the basement rocks (Chain Lakes massif) on the northwest, as well as Lower(?) Ordovician eugeosynclinal rocks on the southeast. The exposures to be visited are mostly along or near the boundaries of the major rock sequences in the southeastern corner of the Chain Lakes quadrangle (see Boone and others, this volume, fig. 1).

The data collected in detailed mapping of the Chain Lakes quadrangle are reviewed on Trip C. They suggest that a major tectonic event involved a block of old continental crust, an ophiolite succession (see Bailey and McCallien, 1960), and associated sediments (flysch). The ophiolite succession here is considered to be composed of coexisting rocks of the alpine complex, pillowed greenstone and metamorphosed quartz latite volcanic rocks (keratophyre) and cherty metasedimentary rocks. This view is compatible with the earlier syntheses by Ells (1887), Boucot (1961), and Boucot and others (1964) who originally proposed that the pre-Silurian rocks in the Boundary Mountains are divisible into two major metamorphic rock sequences on the basis of lithology and age and at least one intrusive series.

Rather different views on the pre-Silurian stratigraphy and structure have been presented in regional syntheses by Cady (1960, 1967), Albee (1961), Marleau (1968), Green and Guidotti (1968), and Harwood (1969) who favor the hypothesis that the pre-Silurian metamorphic rocks constitute a single sequence that variously correlates with rocks along

---

[1]Publication authorized by the Director, U.S. Geological Survey.



Planimetric base from unrectified aerial photographs
(U.S. Geological Survey Series GS-VBKX, 5/24/66)

Geology by E. L. Boudette assisted by G. S. Harwood, Ian Murray, W. B. Thompson, R. J. Thompson, Steven Goldthwaite, and Timothy Targett; 1963–1969

FIGURE C-1. GEOLOGIC MAP OF THE SOUTHEAST CORNER OF THE CHAIN LAKES QUADRANGLE, MAINE, SHOWING ROUTES AND LOCATION OF PRINCIPAL STOPS ON TRIP C

EXPLANATION

UPPER SILURIAN

8

Hardwood Mountain Formation

UNCONFORMITY

UPPER ORDOVICIAN(?)

7

Quartz porphyry dikes and breccia pipes

LOWER(?) ORDOVICIAN

6

Attean Quartz Monzonite

5d | 5c | 5b | 5a | 5a₁ | 5a₂

Rocks of the Alpine complex and related rocks

5d, quartz diorite
5c, massive epidiorite and layered epidiorite quartz diorite
5b, epidiorite - quartz diorite autobreccia
$5a_2$, quartz carbonate veins
$5a_1$, diapiric serpentinite
5a, pyroxenite, gabbro, and massive serpentinite

4

Unnamed Formation

Laminated chloritic metapelite and metagraywacke, and calcareous, chloritic metasandstone and quartzite

3

Unnamed Formation

Dark euxinic metapelite, metagraywacke, and minor feldspathic metasandstone

CAMBRIAN(?) to LOWER ORDOVICIAN(?)

2d | 2c | 2b | 2a

Unnamed Formation

2d, greenstone
2c, metagraywacke and iron formation
2b, quartz latite and keratophyre
2a, greenstone

UNCONFORMITY

PRECAMBRIAN(?)

1b | 1a

Rocks of the [illegible] massif

1b, bedded granofels, [illegible], and gneiss, layered metamorphosed [illegible] rocks and amphibolite
1a, indistinctly layered and massive granofels

Contact, approximately located

U D

Normal
U, upthrown side
D, downthrown side

Reverse thrust
Saw teeth on upper plate

Unclassified

Post-Acadian fault

30		9	78	6
Inclined	Inclined, upright. Ball indicates top known from minor structures	Vertical. Number shows direction of top	Overturned	Inclined, compositional layering in gneisses [illegible]

Bedding and layering

25

Inclined

Foliation parallel to bedding

Foliation

Scheduled stop, Trip

the Maine-New Hampshire boundary which have their equivalents in the Albee, Ammonoosuc, and Partridge Formations (Ordovician) in New Hampshire (Billings, 1956). Both schools of thought are in accord that the pre-Silurian plutonic rocks correlate either with the Highlandcroft Plutonic Series in New Hampshire or with an older series.

The Boundary Mountains anticlinorium appears to culminate in an elongate dome oriented with its long axis on regional strike (see Boone and others, this volume). The core of this dome is underlain by the basement sequence, called for convenience the Chain Lakes massif.[2]

The metamorphic rocks of the massif seen on Stops 1 thru 5 are successively flanked on the southeast (fig. 1) by ophiolite suite rocks of the alpine complex (Stops 4, 5, and 7; units 5a and 5b); the Attean Quartz Monzonite (Albee and Boudette, in press) seen on Stops 8 and 9 (unit 6); greenstone (Stop 10, unit 2a), keratophyre (Stop 11, unit 2b), and cherty rocks of the ophiolite suite; and metamorphosed eugeosynclinal rocks (flysch) (units 3 and 4). The greenstone, keratophyre, and flysch is also referred to as the greenschist sequence here. These pre-Silurian units are unconformably overlain by Silurian and Devonian metasedimentary rocks and intruded by quartz porphyry dikes of Upper Ordovician age, quartz monzonite and granodiorite of the Seven Ponds pluton of Devonian age (Harwood, 1969), and lamprophyre dikes and sills of Triassic(?) age. None of these younger units will be seen on Trip C. Fault systems, probably of at least two widely separated ages, segment the rocks (Stops 4, 5, and 6) and in some places repeat large parts of the section.

Rocks of the Chain Lakes massif include varieties of chloritic and sericitic quartzofeldspathic granofels, gneiss, and schist along with subordinate amphibolite, felsic metavolcanic rocks and vitreous quartzite. Relict to fresh sillimanite, K-feldspar, and other high-grade metamorphic index minerals are locally abundant in rocks of the massif, and anatectic textures are widespread. The granofels and closely associated gneiss are typically massive and are characterized by the presence of lithic fragments and quartz nodules of highly variable size, angularity, and distribution; these are rocks of most unusual texture and structure. The term granofels is applied to identify as much as 75 percent of the massive central part of the massif where no layering is evident. None of the granofels is seen on Trip C. Where rocks of the massif are gneissic or schistose, the foliation is parallel to compositional layering. Relatively abundant small fragments within these layers have a foliation at some angle to the regional foliation.

---

[2]The name Chain Lakes complex has also been used informally, but herein the term complex is restricted to the rocks of the ultramafic series (alpine complex) to avoid confusion.

The granofels and closely associated fragment-bearing gneiss are tentatively assigned to a lowermost stratigraphic subdivision which is of unknown thickness. Metamorphic rocks with conspicuous compositional layering containing few, if any, fragments are correspondingly assigned to an uppermost subdivision about 8,000 feet thick. Both of the subdivisions are upright, face south, and intergrade well, as do subunits within them. Fragment-bearing gneiss (fig. 1, unit 1a) seen on Stops 1 and 2 occurs in the intergradational zone between subunits. Gneiss with good compositional layering is seen at Stops 3, 4, 5, and 6 (unit 1b); this rock is in the lowermost part of the upper subdivision.

No fossils have been found to date the rocks of the Chain Lakes massif, nor has isotopic dating been attempted. Stratigraphic, metamorphic, and lithologic relationships observed in recent work, however, strongly suggest that these rocks are older than Ordovician, probably Precambrian. Certainly they are older than the Middle Ordovician(?) age assigned to the oldest rocks mapped along the Maine-New Hampshire boundary. The evidence for this includes the following:

1. Sedimentary structures show that the rocks of the Chain Lakes massif are the oldest rocks in this area, and that they underlie rocks equivalent to the Albee Formation, Ammonoosuc Volcanics, and Partridge Formation of New Hampshire (Billings, 1956).

2. The unusual fragmented texture of the granofels and gneiss and their association with vitreous quartzite is unknown in northern New Hampshire eugeosynclinal sequences.

3. The mineral composition of fragments and matrix indicate that the dominant source was a nearby probably deeply weathered area from which alumina and silica-rich clastic sediments were derived and mixed with lesser amounts of detritus from a volcanic-igneous terrane. This is at variance with known data on the paleogeography of the Middle Ordovician of northern New England.

4. Rocks of the massif contain two distinctive metamorphic mineral assemblages and two directions of foliation. These metamorphic mineral assemblages and foliations are not recognized in the Middle Ordovician(?) rocks to the west. At Stops 1 and 4, foliated fragments of quartz-feldspar-sillimanite-mica schist occur in chloritic quartz-feldspar-mica sillimanitic gneiss in which the regional foliation is at an angle to foliation in the fragments. This suggests that the fragments were foliated as layered before the metamorphic recrystallization of their host rocks. Furthermore, stress fields accompanying metamorphic events in these rocks have not appreciably changed the initial structural relationship established between host and fragments. At Stop 4 (fig. 2) the intrusion of the alpine complex into gneiss of the massif has produced a contact aureole as much as 400 feet wide which is presently manifested by a recrystallization of the matrix of the gneiss to produce an increase in the amount and grain size of muscovite in the direction of the intrusive

contact. The few fragments observable there indicate that the effect on them was similar. This indicates that fragments and matrix had metamorphically equilibrated before the intrusion of the complex in Early(?) Ordovician time. Because crystallization of the mica appears to be facilitated by the breakdown of sillimanite and K feldspar the contact effect here is regarded as being retrograde. If so, the rocks of the complex must have been prograded before the intrusion of the complex. Because the complex is indicated to be at least Early(?) Ordovician in age such prograde metamorphism could be much older than Early(?) Ordovician, possibly Precambrian.

5. At Stop 3, a greenstone dike correlated with the ophiolite suite has a chlorite-albite-epidote prograde assemblage formed either in Acadian time or during older (Taconic?) metamorphism or both. The older wall rock has a prograde quartz-K feldspar-muscovite-sillimanite assemblage clearly demonstrating the contrast in metamorphic grade and age.

6. An exposure not visited on Trip C, but analogous to that at Stop 4 shows rocks of the south side of the complex in contact with greenstone and metagraywacke (fig. 1, units 2a, 2c, and 3) of the greenschist sequence. Here there is metamorphic amphibole in the greenstone and cordierite-andalusite in the pelitic rocks and some reaction hornfels at the contact. These features are also interpreted as the result of contact metamorphism associated with the intrusion of the complex where the recrystallization was prograde up to the hornblende hornfels facies, according to the index minerals.

7. The rocks of the massif are high-grade metamorphic rocks and the alpine complex rocks along with the contact aureole around them must have been metamorphosed during Acadian time and possibly Taconian as well. Apparently this was the timing of the latest retrograde effects on the metamorphic rocks as well as the latest alteration of rocks of the complex (see also Albee and Boudette, in press) which occurred without profound textural modifications. The interpretation that the relict aureole around the ultramafic complex is, on one hand, impressed on the high-grade metamorphic rocks of the massif, and on the other, impressed upon the low-grade metamorphic rocks of the greenschist sequence, suggests that both sequences existed spatially juxtaposed in essentially their present position before the intrusive event.

The top of the Chain Lakes massif rocks is not exposed in contact with rocks of the greenschist sequence, and no angularity can be demonstrated to show that these units are unconformable. The only evidence of unconformity is the discordance of metamorphic history and the absence of demonstrable faulting at this discordance. Relationships of Ordovician and Cambrian rocks in Vermont and southern New England do not show this discordance of metamorphism, but such a discordance is known at the base of the Paleozoic in western Massachusetts (Norton, 1967) and elsewhere. Consequently, the rocks of the massif are assigned arbitrarily

to the Precambrian(?) in accordance with the interpretation of Ells (1887), Boucot (1961), and Boucot and others (1964). Actually the massif rocks are known only to be Early(?) Ordovician or older; thus a correlation with a somewhat similar sequence of rocks of Early(?) Ordovician or Cambrian(?) age in southern New England is possible (L.R. Page, personal communication, 1969).

Rocks of the greenschist sequence are subdivided in stratigraphic order (fig. 1), from oldest to youngest into basaltic metavolcanic rocks (unit 2a); quartz latite and keratophyre (unit 2b); metagraywacke (unit 2c), dacitic to basaltic metavolcanic rocks (unit 2d), and chert and iron formation (in units 2c and 2d); dark euxinic metapelite and metagraywacke (unit 3); and laminated metapelite and calcareous metasandstone (unit 4). Only the lowermost units 2a and 2b are seen on Stops 10 and 11, respectively.

The greenschist sequence, like the Chain Lakes, is upright and faces southeast, as shown by a combination of minor structural criteria on contacts. This sequence is at least 6,000 feet thick on the northwest, but nearly all the section is repeated on the southeast side of a northeast-trending thrust fault (fig. 1) which dips to the southeast. The entire sequence also appears to stratigraphically thicken to the southeast where it may be in excess of 20,000 feet thick.

The youngest unit (fig. 1, unit 4) in the greenschist sequence, according to Harwood and Berry (1967), grades upward into dated Middle Ordovician rocks in the southeastern part of the Cupsuptic quadrangle. The Attean Quartz Monzonite (fig. 1, unit 6), tentatively dated as Early(?) Ordovician (see below), intrudes the greenschist rocks. Thus, a minimum age spread is implied for the greenschist rocks from Early(?) to Middle Ordovician, but no maximum age can be fixed without knowing the minimum age of the prograde metamorphism in the Chain Lakes rocks. No upper contact of the greenschist sequence is known in this northwestern belt, and the rocks are arbitrarily assigned an Early(?) Ordovician age.

Intrusive rocks almost everywhere separate the Chain Lakes massif on the northwest from the greenschist sequence on the southeast. Near Round Mountain Pond (fig. 1) septa of greenstone (unit 2a) occur near the Chain Lakes, and a contact between greenstone and massif rocks (unit 1a) may be concealed under glacial drift between Round Mountain Pond and Stop 3 where metabasalt float has been found. The intrusive rocks are tentatively subdivided into two series, the older alpine complex (Stops 4, 5, and 7, units 5a thru 5d), and the younger Attean Quartz Monzonite (Stops 9 and 10, unit 6).

The dominant rock of the alpine complex is medium-to coarse-grained epidiorite which is locally massive, but more commonly in lit-par-lit association with quartz diorite of similar texture (Stop 4) or as autobreccia (Stops 4, 5, and 7). Texturally similar quartz diorite

occurs in a discrete body near Round Mountain Pond (fig. 1) where it contains blue quartz (see also Moench and Boudette, this volume, on the lithology of clasts in the Conglomerate of the Rangeley Formation). Pyroxenite, gabbro, and serpentinite with associated rocks are found in subordinate amounts in the basal part of the complex, as shown by detailed mapping in the western part of the Chain Lakes quadrangle.

The mapping to the west has indicated that the base of the complex is relatively enriched in magnesia; the magnesia-rich rocks are overlain in sequence by layered epidiorite-quartz diorite and epidiorite autobreccia, and lastly by massive epidiorite which is cut by the Attean Quartz Monzonite. Diapiric serpentinite with associated quartz-carbonate-fuchsite(?) veins (W.B. Thompson, unpublished report) are found along faults in greenstone (unit 2a) as much as 4 miles away from the serpentinite in the complex. These bodies are believed to be derived from autochthonous counterparts in the complex. None of the diapiric variety of serpentinite will be seen on Trip C.

At Stop 9, Attean Quartz Monzonite along its southeastern boundary presents a textural contrast to that seen to the northwest in the medial part of its pluton at Stop 8. The rock at Stop 9 is much more porphyritic than most and has a discrete aphanitic groundmass, being rather similar to the rocks of the pre-Silurian Rockabema pluton in northern Maine. The texture of the rock at Stop 9 is, in fact, somewhat comparable to that in parts of the quartz latite volcanic rocks seen at Stop 11 about 1,700 feet to the southeast and to that in quartzwacke beds in unit 2c (fig. 1) which are not seen on Trip C. It is important to note that the Attean outcrops at Stops 8 and 9 are structurally beneath the volcanic rocks to the southeast, but younger according to the intrusive relationships.

The alpine complex and the Attean Quartz Monzonite are envisioned as composing an elongate prism of intrusive rock oriented with its long axis on regional strike and with its intermediate axis more or less coincident with the regional dip of the intrusive succession. Layering in rocks of the complex dips 30°-60° SE. (Stop 4). No major repetitions in these rocks are observed, and about 5,000 feet of the complex and 3,000 feet of the Attean Quartz Monzonite is present for a total of 8,000 feet for the plutonic prism as is implied from its average regional dip of about 50°. Both the upper (Stops 4 and 5) and lower contacts (Stops 9 and 10) of the intrusive prism are exposed, but logistics permit only the lower contact to be shown on Trip C. Whether or not the rocks of the complex and Attean Quartz Monzonite are genetically tied, and therefore compose a single series, is presently undetermined. It can only be observed that they bear a close spatial relationship.

The minimum age of the rocks of the complex is fixed by the age of the Attean which intrudes them, and the age of both intrusive series is closely fixed by the Lower(?) Ordovician rocks which they intrude. Radiometric age

work in progress on the Attean Quartz Monzonite (see also Albee and Boudette, in press) indicates that its age is between 450 and 500 m.y. A tentative age assignment of Early(?) Ordovician is made for the Attean, and by implication the rocks of the complex are also Early(?) Ordovician.

At Stop 10, sparse blue quartz clast occur in lapilli beds within the greenstone (unit 2a). The color and texture of the blue quartz clasts are similar to those found in the quartz diorite (unit 5d) near Round Mountain Pond (see p. 11). If the assumption is made that the source of this quartz is in a rock related to that at Round Mountain Pond, then the rocks of the alpine complex and the volcanic rocks of the greenschist belt must have a related genesis. It is on this basis that the rocks of the complex and rocks of the greenschist belt are here regarded as composing an ophiolite succession.

ROAD LOG FOR TRIP C

Assemble at Sarampus Falls State Wayside, 12.8 miles north of the Cathedral Pines Forest Management area, Eustis, Maine, on State Highway #27 at 8:45 AM. Please enter the wayside by the south approach so that vehicles may be turned easily. Unimproved access roads will be followed over most of the trip, and drivers should be prepared for the usual hazards. Special preparations will be made where precautions are required and in the interest of traffic safety on Highway 27. Parking limitations on access roads will limit the trip to no more than 10 cars and about 50 participants. The Chain Lakes and Kennebago Lake topographic quadrangle maps are recommended logistical aids. Copies of the Spencer Lake and Stratton quadrangles would be helpful, but are not essential. Two relatively long traverses into the forest will be taken on Stops 3 and 4; participants should be equipped with appropriate footwear and clothing. Participants should either be equipped with a Brunton compass or plan to traverse close by someone having one in case separation from the group occurs during Stops 3 and 4.

Mileage

0.0 STOP 1. Sarampus Falls State Wayside. Walk across the highway to the outcrop visible on the north. There is a minimum of space between the highway and the outcrop, and the approaches are blind. Please use extreme care. Please do not hammer upon the spectacular fragmented rocks at the south end of the outcrop.

Rocks here are relatively massive, light-to medium-gray, medium- to coarse-grained quartz-feldspar-sillimanite muscovitic and biotitic gneiss containing sparse, but conspicuous, lithic fragments as much as 8 inches across and somewhat smaller quartz nodules. Planar structure (312°, 15° - 30° SW) is shown by light and dark banding believed to be of primary origin. No geometric correspondence exists between the throughgoing gneissic structure and foliation in the lithic fragments. Anatectic features are visible especially on the borders of lithic fragments. Quartz-filled gash veins, several sets of joints, shearing, and incipient brecciation have deformed the rock at Stop 1; these record the latest tectonic event in its history. At least some of these features probably bear a relationship to a fault about 700 feet to the west. A branch of this structure occurs in section in the outcrop on the west side of the road about 800 feet south of Stop 1. The quartz veins here, as well as throughout the entire massif, are sulfide bearing.

The significant features of the rock at this stop are its fragmental texture and its metamorphic mineral assemblage which

records both high-grade (prograde) (sillimanite, anatexis) and low-grade (retrograde) (chlorite, sercitization) events in its history. It is axiomatic that the last event crystallized the chlorite-sericite assemblage in a retrograde alteration. It is important here also to note that the low-angle dip of bedding in the southwest projects these rocks (fig. 1, unit 1a), believed to be upright, beneath those of unit 1b on Bag Pond Mountain.

Return to transportation. Turn left and begin retracing the route southbound on Rte. 27.

0.2 Fault zone (described above) exposed in roadside cut on west. This fault extends more than 3 miles on strike from Bag Pond southeast along the North Branch. It is interpreted as a reverse fault with an appreciable right-lateral component of movement (fig. 1).

1.9 Intersection of Snow Mountain Road. Turn right.

2.4 Bridge over Shadagee Brook.

5.4 STOP 2. Little Alder Stream bridge. Before leaving cars, turn singly (and in order) upon the bridge approach, and carefully back your auto northwest upon the Snow Mountain Road a distance sufficient to accommodate the vehicles that have been following you. Park your auto and walk back to the bridge.

Pavement outcrops in the brook bed are composed of rock of the same medial stratigraphic subunit as the rock at Stop 1. The rock here, though compositionally comparable to that at Stop 1, shows more distinct primary layering and lamination but the bedding is grossly more chaotic. The features preserved here show either the effects of slump with profound cross lamination that was penecontemporaneous with the deposition of the protolith, or an effect of later rheomorphosis during prograde metamorphism. Evidence for the existence of a horizontally directed stress field during prograde metamorphism is lacking at both stops.

Return to transportation and begin retracing the route eastward on Snow Mountain Road.

7.0 STOP 3. Log yard on north where trail departs. Critical exposure to be seen is azimuth 343°, about 2,000 feet on the lowest slopes of Bag Pond Mountain. Access is most easily achieved on lumbering trails familiar to the trip leader. A

moderate pace will be kept because it is desirable to go and return in a group. Please keep together.

At Stop 3, quartzofeldspathic gneiss and metamorphosed laminated, felsic volcanic rocks of unit 1b (fig. 1) are cut by a greenstone (metamorphosed basalt) dike. The dike is oriented on strike about 010°, vertical, and is of variable width to about 3.5 feet. The gneiss and metavolcanic rocks are interrelated, and compositional layering strikes regionally about 070° and dips about 45° to 50° SE. Minor sedimentary structures which persist throughout unit 1b indicate that the sequence is upright and overlies rocks seen at Stop 1 and 2.

Metamorphic considerations outlined for unit 1a at Stops 1 and 2, also apply here (as well as at Stop 4) to unit 1b. The highest metamorphic grade achieved by the greenstone (unit 2a) is indicated to be that of the lower greenschist facies as shown by its chlorite-albite-epidote assemblage. The greenstone (unit 2a) is obviously younger than 1b as shown by the crosscutting relationship, and the textures of unit 2a indicate that it never has contained a mineral assemblage that is typical of any higher metamorphic grade. It is suggested that the protolith of the greenstone was metamorphosed in the second event (retrograde metamorphism) to effect unit 1b when the chlorite-sericite assemblage was formed.

It is considered significant here that the greenstone (unit 2a) is younger than unit 1b.

If the correlation of unit 2a with the Lower(?) Ordovician effusive greenstone is correct (see introductory section, Trip C), contrasting metamorphic histories for units 1b and 2a are indicated in rocks which are here juxtaposed. It is axiomatic that these histories could not have been produced in the two rocks if they have always coexisted. It is interpreted, therefore, on the basis of observations here that unit 1b preceded 2b and travelled a different metamorphic path before acting as host to 2b which probably has intruded into its present position in Early(?) Ordovician time. Thereafter the two units travelled the metamorphic path together. If the protolith of the greenstone is by chance Devonian in age, which is possible, the same observations hold, but the implications do not apply to unit 2a in the southeastern part of the region, and the greenstone here should not be correlated as shown (fig. 1).

Return to vehicles and resume easterly travel.

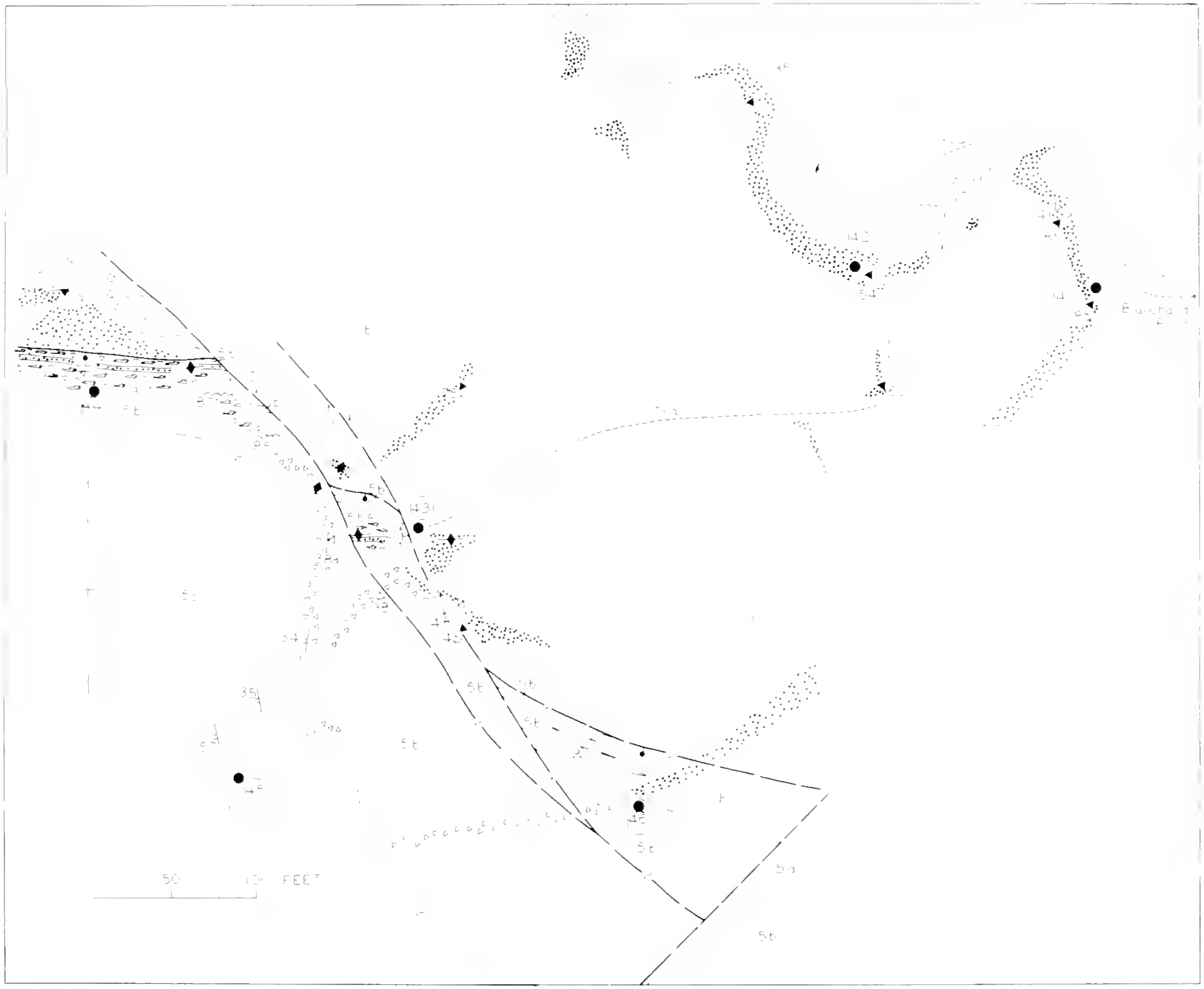
7.9 STOP 4. Blanchard Pond Trail. The exposures to be seen are azimuth 203°, about 2,500 feet from the start of the trail which will be used for access (fig. 1). The same stay-together request outlined for the previous stop is also made here. The route will be a bit soggy in places where beavers have been at work.

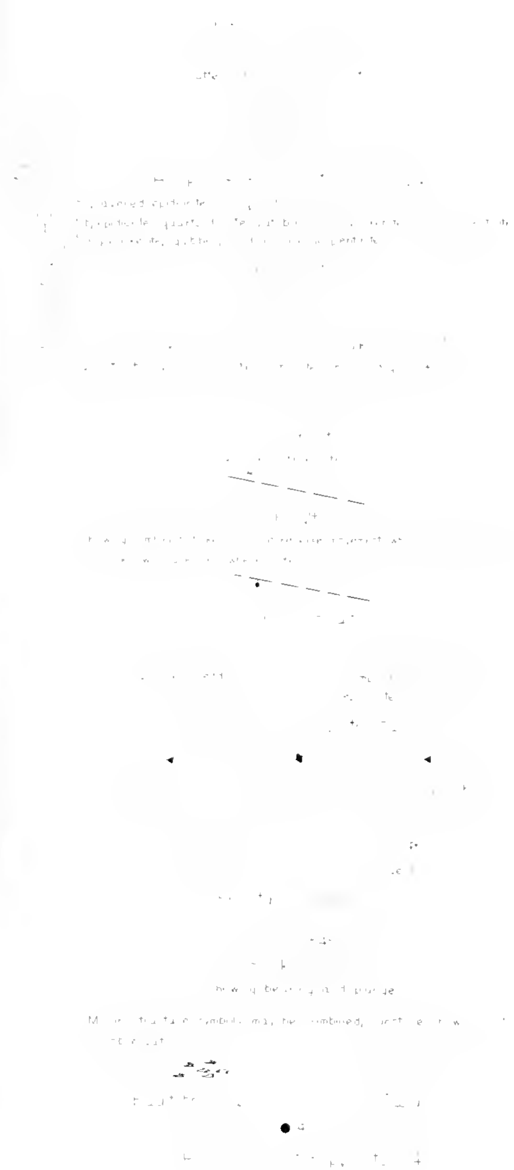
At Stop 4, a series of six principal exposures will be seen in an area about 6,000 feet square (fig. 2) where rocks (unit 1b) along the southern boundary of the Chain Lakes massif are intruded by rocks (units 5b and 5c) of the ultramafic complex. Faults probably of two distinct systems, separable by relative age and geometry, locally segment the rocks of both sequences. Igneous layering in rocks of the complex has a significant bearing upon the regional dip of the sequence and its stratigraphic superposition. Contrasting metamorphic facies comparable to those seen at Stop 3 are also present. Additional metamorphic complications must be considered here, however, such as the diapthoresis and regional retrograde metamorphism of the rocks of the complex and the probable evidence for a relict contact aureole in the intrusive manifested by muscovite textures in unit 1b.

Rocks of the unit 1b (Stops 4.1, 4.2, 4.3, and 4.6) are bedded quartz-feldspar-muscovite biotitic sillimanite gneiss showing good bedding-plane foliation and scattered lithic fragments and quartz nodules (Stop 4.2). Some of the lithic fragments preserve schistosity which is oriented at a large angle to the regional bedding-plane foliation. Muscovite within unit 1b increases both in amount and grain size toward the complex. This feature is preserved in a zone less than 400 feet wide within unit 1b, especially in this area of Stop 4 where faulting displacements appear to be minimal compared with other parts of the southern boundary of unit 1b.

Rocks of the alpine complex are notably altered chloritic varieties of epidiorite, quartz diorite, and pyroxenite with abundant serpentinite and talc segregations. The epidiorite and quartz diorite occur both in a layered sequence and as autobreccia.

Faulting at Stop 4 is typical for the southeastern boundary of the Chain Lakes massif. The oldest faults here are oriented on strike east-west and are either normal faults with the downthrown block on the south or possibly faults with both normal and strike-slip movement. These are offset by a younger fault system composed, in part, of two fault sets oriented on strike to the northwest (figs. 2, 1) and northeast (fig. 1). The





Geology mapped by tape and Brunton method by E.L. Boudette, Steven Southwaite, and Timothy Targett; 1968-69

FIGURE C-3 SCHEMATIC GEOLOGIC MAP OF THE AREA SOUTHWEST OF BLANCHARD POND, CHAIN LAKES QUADRANGLE, MAINE SHOWING ACCESS TRAIL AND SEQUENCE OF OUTCROPS FOR STOP 4

faults in the younger system appear to have compound displacement so that they are reverse, with the northerly block downthrown and right-lateral slip on the northwest-trending set (fig. 2) and left-lateral slip on the northeast-trending set (fig. 1). Some rotation of the older fault system by the younger is possible. Interpretation of slickensides near Stop 4.3 in combination with the mapped horizontal offset and contact of units 1b and 5b (which is assumed to be parallel to the igneous layering) indicates that the younger faults in the northwest set here have each moved the southerly block relatively up about 100 feet on the vertical component and relatively northwest about 100 feet on the horizontal component.

The intrusive contact exposed on the southeast-facing wall (fig. 2, Stop 4.6) is a unique feature in that it is the only known demonstration of the intrusion of rocks of the complex into those of the Chain Lakes sequence. Elsewhere along the boundary, faulting, which postdates the emplacement of the complex, predominates and obscures the intrusive relationships. At Stop 4.6, the intrusive relationship clearly demonstrates that the Chain Lakes is older. This observation, obviously, has important regional implications (see introductory section, Trip C) and is the basis for explaining the gradational nature of the muscovite texture in unit 1b here in terms of a relict contact aureole (see p. 7). It is axiomatic that the age of this aureole, if the interpretation is correct, must be that of the rocks of the complex (Early(?) Ordovician). Because this aureole is interpreted to have been impressed upon the Chain Lakes rocks while in their highest prograde metamorphic facies, it is implied that unit 1b predated the rocks of the complex in its present textural form as a gneiss. A situation analogous to that at Stop 3 is, therefore, suggested here.

Return to transportation and resume travel.

8.9 Turn right and proceed south on Rte. 27.

9.2 STOP 5. (Jim Pond/Alder Stream township line). Park off the highway as directed and walk south on highway to exposure.

Please use the safety precautions outlined for Stop 1.

Note: Road construction at Stop 5 began in fall 1969. At the time that the road log was prepared it was not known whether the outcrops would be improved or obscured by this work.

Cataclastic quartz-feldspar-muscovite chloritic gneiss of the Chain Lakes massif on the northwest is probably in fault contact

with epidiorite autobreccia, altered pyroxenite, and serpentinite of the ultramafic complex on the southeast (fig. 1). The rocks of the massif seen here are probably among the lowermost layers of the pluton as at Stop 4 because of the presence of pyroxenite and serpentinite.

Return to transportation.

10.5 Roadside outcrops of sheared Attean Quartz Monzonite (unit 6).

11.0 Roadside outcrops of Lower(?) Ordovician greenstone (unit 2a).

11.8 Roadcut (unit 2d) in sheared, serpentinized greenstone containing jasper lenses and quartz-calcite veins. The quarry is a source of highway riprap.

12.1 Turn left on Jim Pond Camps access road.

12.3 Bridge over North Branch Dead River. Bear left at intersection onto lumbering road.

12.9 Intersection; bear left.

16.4 Turn around as directed where the road is wide and its shoulders are dry and level. Begin retracing route in the southerly direction.

16.6 STOP 6. Foliated granofels of the Chain Lakes massif.

The granofels at Stop 6 is lithologically comparable to that seen in the previous two stops. Here, however, foliation, oriented 060°, 75° SE, is strongly developed and conspicuous. This foliation is believed attributable to shearing related to either normal or strike-slip local faulting which gradually cuts out all the mafic rocks in the complex to the east until the Attean Quartz Monzonite is in direct contact with the massif on the east side of the Chain Lakes quadrangle (fig. 1).

17.1 Viles Pond Brook

17.2 STOP 7. Epidiorite-quartz diorite autobreccia (unit 5b) of the ultramafic complex.

Rocks here are comparable to some of those seen at Stops 4 and 5, but evidence for serpentinization is inconspicuous or serpentinite is lacking entirely. The outcrop is a particularly good 3-D exposure of the unit. The epidiorite, as the abundant facies, is composed essentially of chloritized amphibole

and plagioclase. The epidiorite is cut by a boxwork of finer-grained quartz diorite veins and, rarely, quartz veins. Rare xenoliths of granofels of the complex may be found in this unit which attests to the relative age of the sequences as does the contact at Stop 4.6. The autobreccia lacks conspicuous deformation here, but it is cut by several sets of fractures tending to brecciate parts of the unit.

17.7 Roadside outcrop of epidiorite autobreccia.

18.2 Intrusive contact of the Attean Quartz Monzonite with the alpine complex is concealed beneath Pleistocene ground moraine.

18.3 STOP 8. Attean Quartz Monzonite.

Attean Quartz Monzonite crops out here as part of a body about 3,000 feet thick (fig. 1). The rock seen is composed of about equal amounts of quartz, altered plagioclase, and potassium feldspar with 5 percent or less of altered mafic accessory minerals; it is sheared, foliated, and chlorite and epidote are conspicuous alteration products.

18.8 STOP 9. Attean Quartz Monzonite.

Rock on the southeastern side of the pluton (fig. 1) presents a textural contrast to that seen at the previous stop. Quartz and some feldspar in the Attean here is of the same grain size, but the bulk of the feldspar is more altered and occurs as groundmass components. The Attean here is indicated to be structurally beneath the volcanic rocks to the south by graded beds and pillow-top observations, and its intrusive relationship with unit 2a is unequivocally established.

19.0 STOP 10. Lower(?) Ordovician greenstone with lapilli beds.

The greenstone here (unit 2a) is composed of amygdaloidal metabasalt with poorly developed pillow structure which is interlayered with graywacke beds as much as 1 foot thick composed principally of lapilli, feldspar, and quartz. Pink to brownish calcite composes the amygdules, and rare blue quartz clasts occur in the lapilli beds. (Please do not hammer upon, or remove these quartz clasts.) Layering in this unit strikes 043° and dips about 78° NW. Grading and crosslamination in the lapilli beds indicates that the layering is slightly overturned; the regional sequence faces southeast. The metamorphic grade is indicated to be in the lower greenschist facies by a chlorite-albite-epidote assemblage; no textures suggest that it was ever in a facies characterized by higher temperature.

19.2 STOP 11. Metamorphosed quartz latite volcanic rocks of the Lower(?) Ordovician sequence.

The greenstone at the last stop is stratigraphically overlain by metamorphosed lava (keratophyre), breccia, and ash flow of quartz latite composition (unit 2b); the contact is sharply defined. The sequence of outcrops here represents nearly two-thirds of an unbroken section of felsic volcanic rocks which is about 2,000 feet thick. These volcanic rocks are part of a lenticle that extends on regional strike from the northwestern ninth of Kennebago quadrangle through Stop 11 to the northeast for an unknown distance.

Here these volcanic rocks appear in perhaps their most spectacular accumulation. They are part of an unbroken southeast-facing unit which, in contrast to the lower boundary, grades well into overlying metagraywacke-iron formation and greenstone units (units 2c and 2d). Interbeds of ferruginous chert and quartz-rich graywacke increase in abundance toward the top of the unit and finally predominate (unit 2c).

In unit 2c, thick quartz-rich graywacke beds are commonly associated with the iron formation. These are not exposed in the section here but are observable in the ridges on strike.

The metamorphic grade of rocks at Stop 11 is compatible with that at Stop 10, and indications are that rocks at both stops have had the same history of pressure-temperature since deposition.

Return to transportation, remain on principal road, and retrace the route back to Rte. 27.

Note: The following notes are added in the interest of continuity. Refer to Boone and others (this volume, fig. 1) for geology. Unscheduled stops may be added, if interest and time permit, to elucidate the stratigraphic sequence in Lower(?) Ordovician rocks.

20.7 Intersection with Rte. 27; turn left and proceed south.

20.8 Highway bridge over Alder Stream.

21.2 Intersection of Alder Stream Road at Alder Stream State Campsite.

21.3 Roadside outcrop on right is phyllite and quartzite of unnamed formation (unit 4).

21.4 Trace of major thrust fault concealed beneath ground moraine which repeats the entire Lower(?) Ordovician sequence to the southeast and forms the southeastern boundary of the Moose River synclinorium to the northeast (Boucot, 1961). Outcrops of greenstone

begin 0.1 miles south of here and continue for 0.7 miles. These are part of unit 2a, much thickened, seen at Stop 10. Pillow lavas are quite common in this belt of greenstone and locally serpentiniized zones.

22.5 Greenbush Pond visible on left.

22.6 Greenbush State Campsite.

23.6 Roadside outcrops of greenstone are continuous for 0.5 mile. These are in the same unit with the greenstone southeast of the fault.

24.5 Intersection of "CCC" Road on the left.

24.7 Jim Pond Township/Eustis Town line.

24.9 Intersection of East Tea Pond Road.

25.3 Intersection of Jim Pond Road.

25.4 Roadside outcrops of chloritic phyllite of unit 4 are continuous for 0.3 miles to south.

26.0 Eustis Village.

Note: The highway south traverses Pleistocene valley train deposits; features include an esker with a kettle hole to the west and deltaic and glaciolacustrine deposits further south.

28.4 Cathedral Pines Forest Management Area -- End of Trip C.

Note: Participants in Trip C who are proceeding to coastal areas, Boston, or the southern Interstate Highway system will find Rte. 27 to Farmington, Maine more direct than traveling via Rangeley.

LITERATURE CITED

Albee, A. L., 1961, Boundary Mountain [sic] anticlinorium, west-central Maine and northern New Hampshire: U.S. Geol. Survey Prof. Paper 424-C, p. C51-C54.

_____ and Boudette, E. L., 1971, Geology of the Attean quadrangle, Maine: U.S. Geol. Survey Bull. 1297 (in press).

Bailey, E. B., and McCallien, W. J., 1960, Some aspects of the Steinmann trinity, mainly chemical: Geol. Soc. London Quart. Jour., v. 116, pt. 4, p. 365-395.

Billings, M. P., 1956, Bedrock geology, pt. 2 of The geology of New Hampshire: Concord, N.H., New Hampshire State Plan. and Devel. Comm., 203 p.

Boucot, A. J., 1961, Stratigraphy of the Moose River synclinorium, Maine: U.S. Geol. Survey Bull. 1111-E., p. 153-188.

_____ Griscom, Andrew, and Allingham, J. W., 1964, Geologic and aeromagnetic map of northern Maine: U.S. Geol. Survey Geophys. Inv. Map GP-312.

Cady, W. M., 1960, Stratigraphic and geotectonic relations in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, no. 5, p. 531-576.

_____ 1967, Geosynclinal setting of the Appalachian Mountains in southeastern Quebec and northwestern New England, in Clark, T. H., ed., Appalachian tectonics: Royal Soc. Canada, Spec. Pub. No. 10, p. 57-68.

Ells, R. W., 1887, Report on the geology of a portion of the Eastern Townships [Quebec], relating more especially to the counties of Compton, Stanstead, Beauce, Richmond, and Wolfe: Canada Geol. Nat. History Survey, 2d Ann. Rept., 1886, pt. J., 70 p., map.

Green, J. C., and Guidotti, C. V., 1968, Geology of the Boundary Mountain anticlinorium in northern New Hampshire and northwestern Maine, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., Jr., eds., Studies of Appalachian geology--northern and maritime: New York, Interscience, p. 225-266.

Harwood, D. S., 1969, The Second Lake anticline--A major structure on the northwest limb of the Boundary Mountain [sic] anticlinorium, northern New Hampshire, west-central Maine, and adjacent Quebec: U.S. Geol. Survey Prof. Paper 650-D, p. D106-D115.

Harwood, D. S., and Berry, W. B. N., 1967, Fossiliferous lower Paleozoic rocks in the Cupsuptic quadrangle, west-central Maine: U.S. Geol. Survey Prof. Paper 575-D, p. D16-D23.

Marleau, R. A., 1968, Woburn-East Megantic-Armstrong area, Frontenac and Beauce Counties: Quebec Dept. Nat. Resources Geol. Rept. 131, 55 p.

Norton, S. A., 1967, Geology of the Windsor quadrangle, Massachusetts: U.S. Geol. Survey Open-File Rept., 210 p.

TRIP D

Evidence for premetamorphic faulting in the Rangeley quadrangle, western Maine[1]

by

R. H. Moench
U.S. Geological Survey
Denver, Colorado

The purpose of this trip is to examine the Hill 2808 fault, one of three major premetamorphic faults that have been delineated by recent mapping in the Phillips, Rangeley, and Rumford quadrangles (Moench, 1969; Moench, 1970 a, b). For detailed descriptions and interpretations, of the faults, participants are referred to Moench (1970, a). In brief, each fault is considered to be a normal fault, downthrown on the southeast toward the center of the Merrimack sedimentary trough. Each fault, the large syncline on its southeast side, and the major anticline farther southeast are interpreted to constitute a fault-fold slump unit, not unlike the much gentler pattern of down-to-basin faults and their associated folds of the Gulf Coast (Cloos, 1968). The Hill 2808 fault, Mountain Pond syncline, and Brimstone Mountain anticline (fig. 1), represent the best exposed fault-fold unit in the Rangeley area. Figure D-2 illustrates my interpretation of the fault-fold unit in three dimensions. The fault is interpreted as a normal hinge fault along which displacement (down on the southeast) increases toward the southwest in the direction of plunge of the anticline. Increasing vertical structural relief in the syncline-anticline pair appears, thus, to be related to increasing normal displacement along the fault.

Stops D-1A and D-1B are on the approximate hinge of the Hill 2808 fault where parts C and B of the Rangeley Formation begin to converge against [the lower] part A (fig. 1). The stop is in the northwest limb of the Mountain Pond syncline. Critical features here are the southwest plunges of small folds and lineations, defined by intersections between bedding and slaty cleavage or schistosity. In contrast, folds and lineations on the nose of the Brimstone Mountain anticline to the southeast (fig. 1) plunge in the opposite direction. Because these features of opposite plunge share the same axial surface schistosity, they represent the same deformations. Moreover, they represent the oldest and most important deformation in the region.

---

[1]Publication authorized by the Director, U.S. Geological Survey

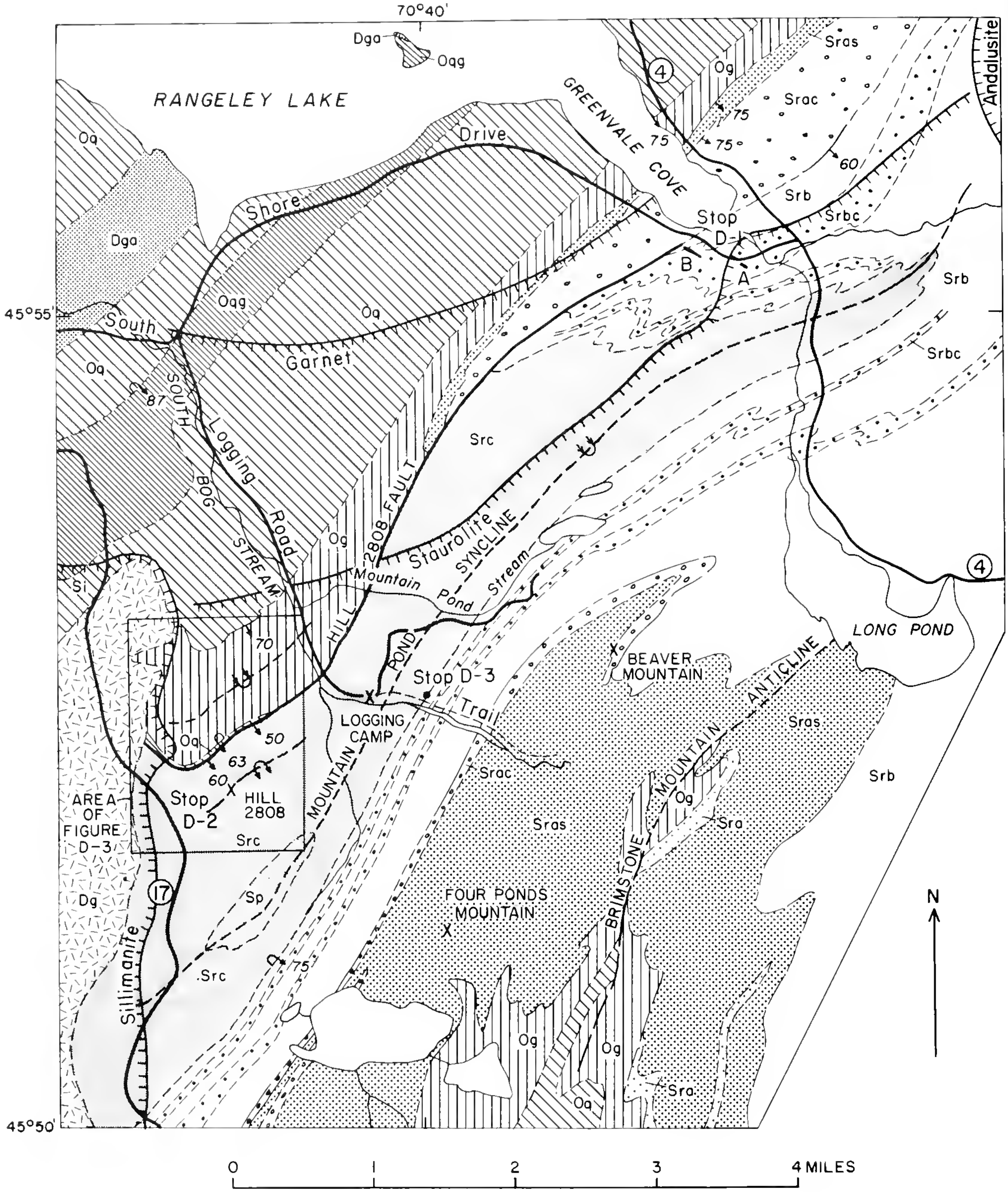


Figure 1.--Geologic map of area of Trip D. (From Moench, 1970 b)

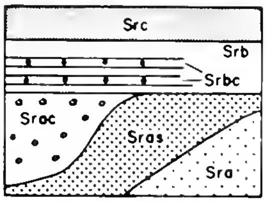
EXPLANATION

Granitic rocks

Gabbroic rocks

DEVONIAN

Perry Mountain Formation



Rangeley Formation

Part C:

Src, *dominantly metashale; subordinate quartz metaconglomerate*

Part B:

Srb, *dominantly metashale*

Srbc, *metashale and quartz-rich polymictic metaconglomerate*

Part A:

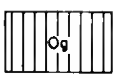
Srac, *polymictic metaconglomerate*

Sras, *arkosic metasandstone*

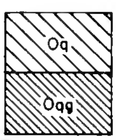
Sra, *dominantly metashale*

Lower Silurian(?)

SILURIAN(?)



Greenvale Cove Formation



Quimby Formation

Oq, *dominantly metashale*

Oqg, *dominantly metashale and felsic metavolcanic rocks*

Upper Ordovician(?)

ORDOVICIAN(?)

63 Overturned

75 Right-side-up

Contact

Showing dip where exposed. Dashed where approximately located; short dashed where gradational or inferred in areas of no outcrop

60

Fault

Showing dip where exposed. Dashed where approximately located; short dashed where inferred

Anticline

Syncline

Overturned folds

Dashed where approximately located; short dashed where inferred

Metamorphic isograd

Hachures on side of greater intensity

Stop D-3

A B

Stop D-1

Field trip stops

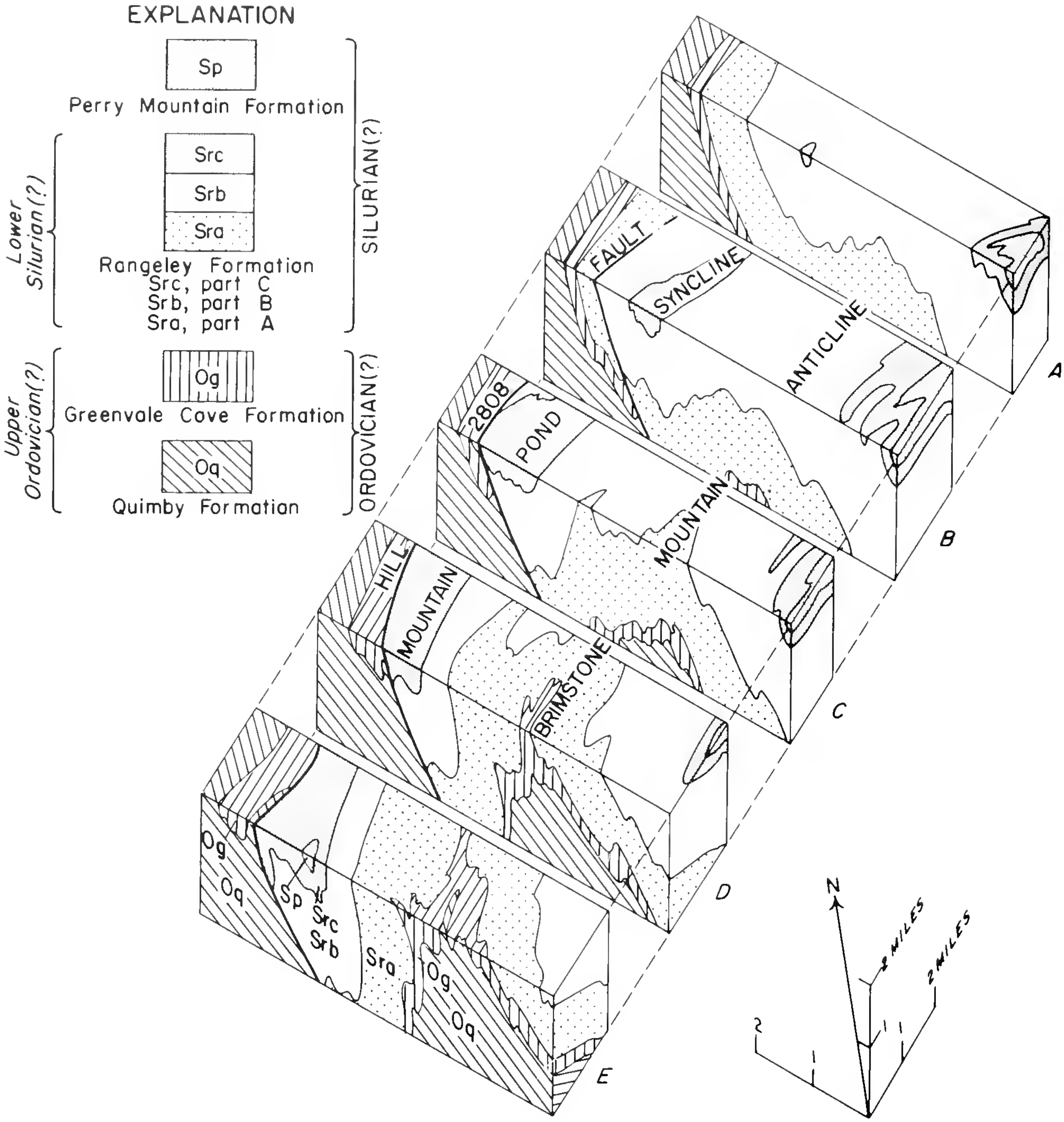


Figure 2.--Block diagram showing relations of Mountain Pond syncline and Brimstone Mountain anticline to Hill 2808 fault. (From Moench, 1970 a, fig.10)

Stop 2, along the crest of Hill 2808 (figs. 1, 3), crosses the area where the Hill 2808 fault is demonstrated as a major structural feature and is exposed in several ouucrops (fig. 3, outcrops A-F). Critical features here are: (1) tops of bedding determined from sedimentary features, which indicate that younger rocks on the southeast face toward older rocks on the northwest, and (2) characteristics of the fault contact that demonstrate the precleavage age of the fault, its southeast dip, and the plastic, semiconsolidated character of the wall materials during faulting.

If time and logistics permit, stop D-3 will be visited. Here, metamorphosed clastic dikes that are parallel or nearly parallel to the oldest generation of schistosity are well displayed (Moench, 1966, figs. 4, 5; pl. 2). These dikes and others in lower grade slates are the principal evidence for the origin of slaty cleavage by tectonic compaction (Maxwell, 1962). (See also Braddock, 1970; Powell, 1969.)

Participants should be prepared for a steep 700-foot climb to the top of Hill 2808 at stop D-2. Critical outcrops are on the steep slope of the ridge; blowdown, slash, and rather dangerous loose talus will be avoided as much as possible.

I am grateful to E. L. Boudette, H. R. Dixon, and R. W. Schnabel for constructive comments on the manuscript.

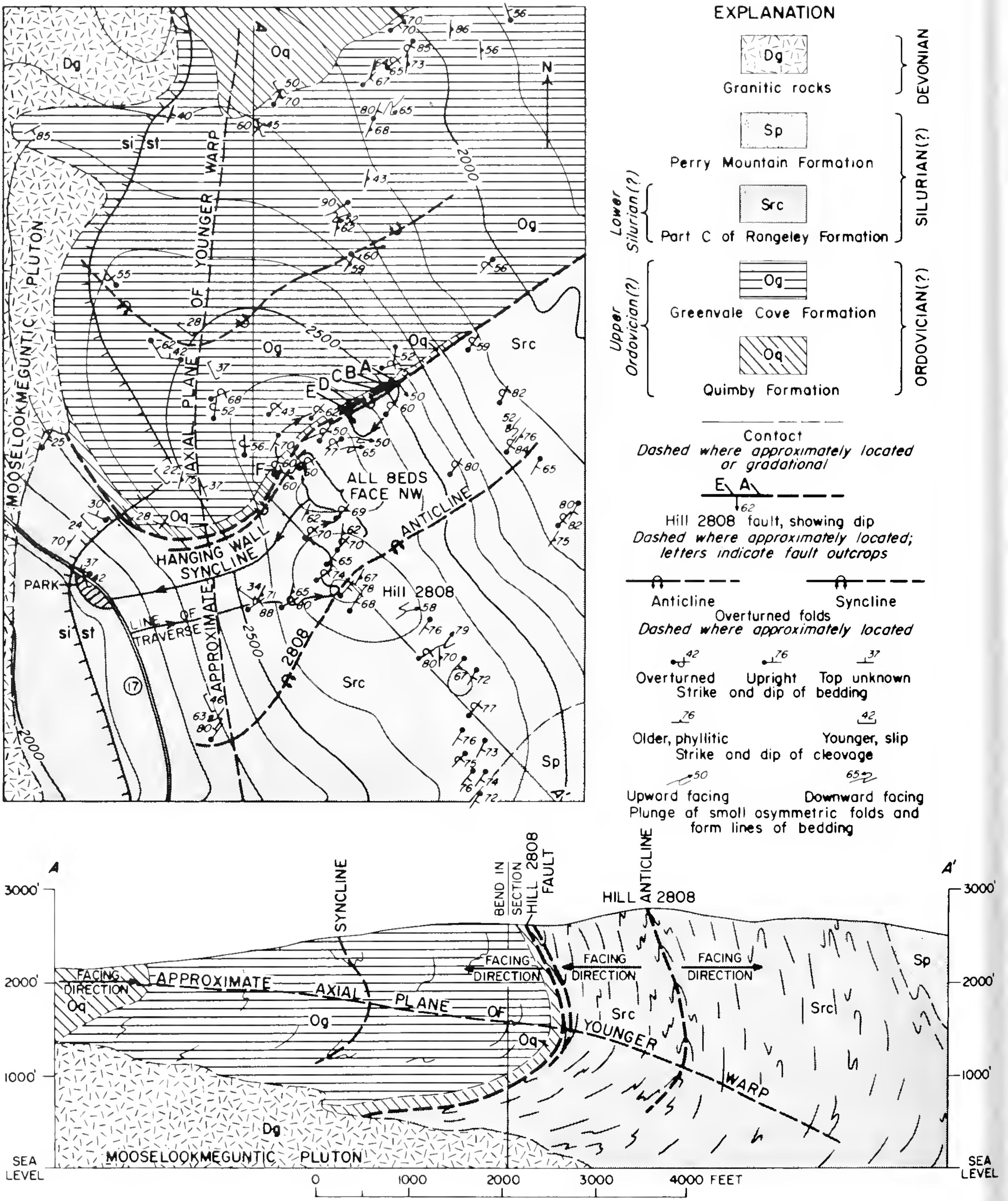


Figure 3.--Geologic map and section of area of stop D-2. (Modified from Moench, 1970 a,fig. 6)

LOG FOR TRIP D

Assemble at Rangeley Chamber of Commerce building. Owing to small outcrops and limited parking space, participation must be limited to 50 people and transportation facilities to 10 automobiles.

Mileage

0.0 Rangeley Chamber of Commerce building. Drive southeast on Route 4.

2.3-2.8 Large outcrops on left of Greenvale Cove Formation and part A of Rangeley Formation (conglomerate) described under trip A-1.

3.4 Turn right onto South Shore Drive (to Rangeley Lake State Park) and cross inlet to Rangeley Lake.

3.7 STOP 1A. Park near abandoned yellow house on left and walk to outcrops south of house and gray tar-paper garage.

Rocks are typical of part B of Rangeley Formation. Gray pelitic phyllite (metashale) with pseudomorphs chlorite and sericite after staurolite is dominant. Phyllite is irregularly interbedded with conglomeratic rocks and metasandstone showing graded bedding. Attitude is N. 75° E., 55° S., right side up; bedding-cleavage intersections plunge 30°, S. 50° W.

Proceed about 175 feet S. 75° W. to small knoll. Folds are exposed that plunge 10°-35°, S. 52° W.

Return to cars.

3.8 Low outcrop on left of metashales, pebbly metamudstone, and thin beds of metasandstone. Younger rocks are to the southeast; bedding-cleavage intersections plunge southwest. Pseudomorphs of chlorite and sericite after staurolite are small and sparse.

3.9 Low outcrop on left of gray metashale and pebbly metashale. Pseudomorphs after staurolite are absent.

4.1 STOP 1B. Park near white house on left ("The Green Bough"); walk up driveway and follow inconspicuous trail into woods south of the house.

Starting at the first outcrop along the trail, walk northwest along the contour across large outcrops of typical pelitic to conglomeratic rocks of part B of the Rangeley. Typical structural styles of the northwest limb of the Mountain Pond syncline

are displayed here. Bedding strikes N. 65°-90° E., dips south, and is right side up; bedding-cleavage intersections plunge southwest. Small metamorphosed clastic dikes along phyllitic cleavage are present in this sequence.

Conglomerate of part A is well exposed in the woods 300-500 feet northwest of the house, and along the shore of Greenvale Cove to the north. Attitudes are about N. 50°-60° E., 80° SE, right side up.

The Hill 2808 fault was not seen in outcrop here, but probably follows the boundary between the conglomeratic part A and dominantly pelitic part B of the Rangeley. Evidence for the fault at this locality is: (1) the abrupt thinning of the lowermost pelitic zone (typically 1,000-1,500 ft. thick where complete) of part B against metaconglomerates of part A; and (2) the convergence of bedding. In part B on the southeast, bedding strikes N. 65°-90° E., and dips south, whereas in the coarse clastics of part A on the northwest, bedding strikes N. 50°-60° E., and dips steeply southeast.

4.3 Outcrop on left at summer cabin ("Dunwirk"); massive feldspathic metasandstone.

4.6 Low outcrops on left and in brook; graded beds of massive feldspathic metasandstone and sparse metaconglomerate. Attitude is N. 40°-45° E., vertical; younger rocks are to the southeast.

5.2 Small outcrop on right of Quimby Formation. Rock is dark-gray rusty-weathering sulfidic slate with thin beds of fine-grained metagraywacke. Slip cleavage strikes northwest and dips steeply northeast.

8.3 Pass entrance to Rangeley Lake State Park (on right).

8.6 Pass junction of logging road (on left).

8.8 Bridge over South Bog Stream.

11.0 Turn left onto route 17.

12.3-12.6 Low outcrops on right, rusty-weathering Quimby metagraywacke.

12.6 Outlook over Rangeley Lake.

12.9-13.2 Scattered outcrops on right of Quimby metagraywacke.

16.2 STOP D-2. Park in turnout on right; set altimeter at 2,130 feet; walk 500 feet southwest along road; walk left into woods on bearing N. 80° E. (See fig. D-3.)

Elevation 2,550 feet; lowest outcrops. Rocks from here to summit are irregularly interbedded metashale, feldspathic quartzite, and quartz metaconglomerate of part C of Rangeley Formation. Beds strike northeast, dip steeply southeast, and most are overturned to the northwest.

In the lower outcrops, porphyroblasts of staurolite are fresh, sharply euhedral, and average about 2 mm in length. They have grown across a conspicuous slip cleavage that strikes about N. 55° W., and dips about 35° N. In Higher outcrops, staurolite is commonly altered to chlorite and sericite.

Summit of Hill 2808; extensive pavement outcrops displaying typical rock types of part C of Rangeley Formation. Rocks are tightly folded,with alternating top directions on both sides of the crestline of the Hill 2808 anticline (fig. D-3). Fold axes are subhorizontal or plunge gently southwest. Fine-grained schistosity is slightly diagonal to the axial surface of a tight syncline.

Traverse about north along ridge from summit, and keep left of worst blowdown.

Between the summit and the first small knoll (not shown on fig. D-3), about 400 feet to the north, facing directions alternate several times from northwest to southeast.

Between the knoll and the trough line of the hanging wall syncline near the Hill 2808 fault (fig. D-3)-- a horizontal distance across strike of about 900 feet--facing directions are consistently northwest. Exceptions are two small folds that were found on the east side of the ridge (fig. D-3). The trough line of the hanging-wall syncline is 0-75 feet from the contact between part C of the Rangeley Formation and the Quimby Formation, and is apparently truncated just downhill from outcrop A (fig. D-3).

Because tops of beds within at least 900 feet of part C of the Rangeley face northwest toward older rocks of the Quimby, the contact must be a fault. Obviously, this 900 feet of section, plus the underlying parts B and A of the Rangeley Formation and the Greenvale Cove Formation, cannot be accommodated within the 0- to 75-foot-wide northwest limb of the hanging-wall syncline. On the assumption that original stratigraphic thicknesses change only slightly subparallel to the regional strike, about 9,500 feet of section is missing along the fault.

Outcrop F of the Hill 2808 fault (fig. D-3); rocks characteristic of the uppermost part of the Quimby Formation are exposed on the north. They are dark-gray carbonaceous rusty-weathering phyllitic metashale

interbedded with thin beds of metasandstone and calc-silicate rocks. Tops of poorly preserved graded beds apparently face northwest. Rocks on the south are metasandstone and subordinate amounts of quartz conglomerate and metashale of part C of the Rangeley. Tops of beds in the Rangeley face southeast; bedding is irregularly folded at the south end of the outcrop and is truncated at a low angle against the contact. The fault zone is represented by a 2-foot-thick zone of unbedded carbonaceous phyllite, probably derived from the Quimby. The hanging wall is sharp. The footwall is indefinite and is defined only by the first appearance of bedding in the Quimby. Fine-grained schistosity on both sides of the fault strikes uniformly N. 10°-20°E., dips about 65° E. (slightly steeper than the hanging wall), and has not been deformed by the fault.

Traverse north across outcrops of laminated nonrusty fine-grained clastics of the Greenvale Cove Formation to outcrops E-A of the Hill 2808 fault on the east side of the ridge (fig. D-3). The hanging wall is commonly sharp and smooth, but locally gives way to an extremely irregular boundary marked by irregular protrusions of Rangeley metasandstone into Quimby-derived phyllite of the fault zone. Note the absence of breccia or mylonite. Fine-grained schistosity is uniform in attitude and crosses the faults at an angle that is slightly steeper than the dip of the fault.

Part C of the Rangeley Formation is exceptionally well exposed in a belt 300 feet wide across strike at outcrop E. Except for the narrow northwest limb of the hanging-wall syncline, the tops of all beds within this belt face northwest. Access to these outcrops is dangerous, however, owing to active talus and loose vegetation on the steep slope.

Return to automobiles.

STOP 3. (fig. 1).-- Search for the pertinent outcrop is not recommended to those unaccompanied by the author. The outcrop is in an area of dense slash and blowdown several tens of feet off a thickly overgrown logging trail now used only by a large ferocious bull moose. The following directions are provided for those who are willing to hazard both the moose and an unsuccessful search. The outcrop should not be difficult to find when deciduous vegetation is bare in May or October.

Drive south on the logging road that intersects South Shore Drive about 0.2 mile east of the bridge over South Bog Stream (fig. D-1).

Mileage

0.0 Junction South Shore Drive and logging road.

0.7-1.5	Gravel pits on left.
2.0 (approx.)	Logging road crosses Mountain Pond Stream. You may have to walk from here.
1.0 (approx.)	Abandoned logging camp; road bends from east to northeast.
1.1	Principal logging road bends to north, and inconspicuous abandoned swampy logging trail branches to the east. The trail approximately follows the boundary between uncut timber on the south and the worst sort of slash and blowdown on the north. Pace approximately 1,750 feet east and southeast along the trail. The pertinent outcrop is in the slash about 100-200 feet north of the trail. It is an isolated, conspicuous, apparently frost-heaved block that stands 10-12 feet above the ground. Horizontal dimensions are about 15 by 25 feet. Other less conspicuous outcrops and rubble, all characteristic rock types of part C of the Rangeley Formation, are present nearby. Although apparently frost-heaved, the block is not significantly disoriented, for the schistosity has the same orientation as it has in nearby outcrops.

The lower three quarters of the block is composed of gray pelitic phyllite (metashale) with abundant pseudomorphs of chlorite and sericite after staurolite. The upper quarter is a thick-graded bed of pale-orange or light-tan garnetiferous biotite-plagioclase-quartz granofels (metasandstone). Bedding is nearly flat-lying, and defines the trough of a syncline that plunges gently northeast. Axial surface schistosity strikes N. 20°-30° E., dips steeply, and refracts in the normal manner where it crosses beds of different composition. Two generations of younger slip cleavage are recognizable as well: (1) rather flat-lying slip cleavage with subparallel tabular porphyroblasts of chlorite; (2) a more conspicuous slip cleavage and parting that dips about 70° NW.

Several metamorphosed clastic dikes are conspicuous on the southwest face of the outcrop (Moench, 1966, figs. 4,5; pl. 2). The dikes are thinly tabular, of various sizes, and are parallel, or nearly parallel, to the axial surface of the fold and to the schistosity, which is the oldest generation of cleavage in the rock. Tabular dikes of metasandstone extend downward from their source beds along the schistosity. Some metasandstone dikes are paired with dikes of metashale that extend upward from their source beds along refracted schistosity in metasandstone.

REFERENCES CITED

Braddock, W. A., 1970, The origin of slaty cleavage--evidence from Precambrian rocks in Colorado: Geol. Soc. America Bull., v. 81, no. 2, p. 589-600.

Cloos, Ernst, 1968, Experimental analysis of Gulf Coast fracture patterns: Am. Assoc. Petroleum Geologists Bull., v. 52, no. 3, p. 420-444.

Maxwell, J. C., 1962, Origin of slaty and fracture cleavage in the Delaware Water Gap area, New Jersey and Pennsylvania, in Engel, A. E. J., James, H. L., and Leonard, B. F., eds., Petrologic studies--(Buddington volume): Geol. Soc. America, P. 281-311.

Moench, R. H., 1966, Relation of S_2 schistosity to metamorphosed clastic dikes, Rangeley-Phillips area, Maine: Geol. Soc. America Bull., v. 77, no. 12, p. 1449-1462.

_____ 1969, Premetamorphic down-to-basin faulting and folding in the Rangeley area, western Maine (abs.), in Program with Abstracts for 1969: Geol. Soc. America Ann. Mtg., Atlantic City, New Jersey, p. 152.

_____ 1970a, Premetamorphic down-to-basin faulting, folding and tectonic dewatering, Rangeley area, western Maine: Geol. Soc. America Bull. v. 81, p. 1463-1496.

_____ 1970b, Geologic map of the Rangeley and Phillips quadrangle, Maine: U.S. Geol. Survey Misc. Geol. Inv. Map I-605 (in press).

Powell, C. McA., 1969, Intrusive sandstone dykes in the Siamo Slate near Negaunee, Michigan: Geol. Soc. America Bull., v. 80, no. 12, p. 2585-2594.

Pleistocene History and Glacio-tectonic Features in the Lac Mégantic Region, Québec[1]

by

William W. Shilts
Geological Survey of Canada

The road log which follows provides detailed descriptions of field localities which define the glacial stratigraphy of the Lac-Mégantic area. Evidence for three glaciations separated by two nonglacial intervals includes three stratigraphically distinct tills, lacustrine deposits related to Glacial Lake Gayhurst, a complex sequence of lake sediments and fine gravels (Massawippi Formation), late-glacial recessional moraines, and striae indicating at least two different ice-flow directions.

In addition, large-scale slump blocks involving thick sections of lacustrine sediments and till, and complex deformation structures in ice-contact sediments will be examined.

---

[1]Publication authorized by the Director, Geological Survey of Canada.

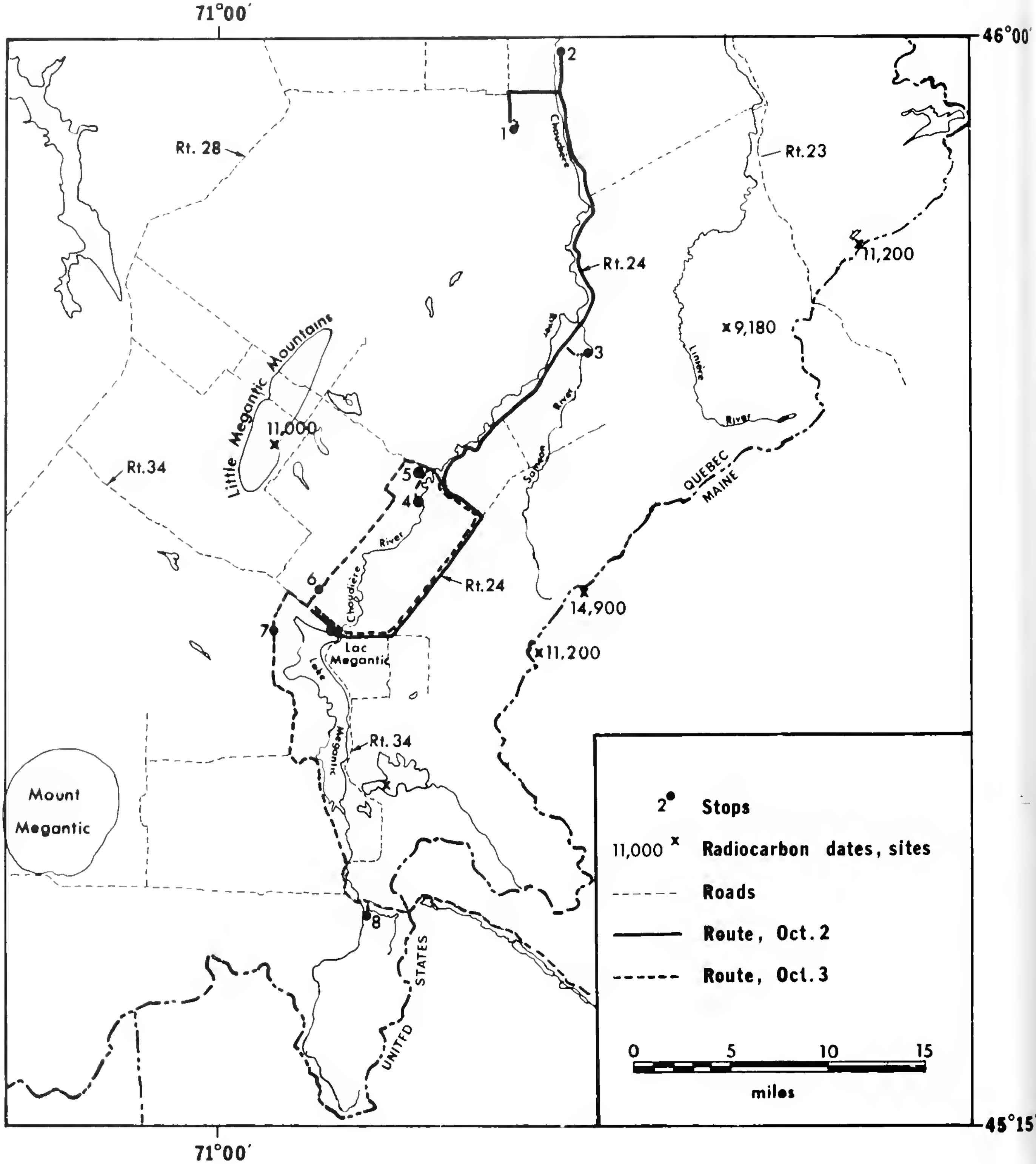

71°00′
46°00′
45°15′
Rt. 28
Rt.23
Rt.24
Rt.34
Rt.24
Rt.34
Chaudière
Little Megantic Mountains
11,000
11,200
9,180
14,900
11,200
Linière
River
Samson
River
QUEBEC
MAINE
Choudière
River
Lac Megantic
Lake
Megantic
Mount Megantic
UNITED
STATES
1
2
3
4
5
6
7
8
Stops
11,000
Radiocarbon dates, sites
Roads
Route, Oct. 2
Route, Oct. 3
0
5
10
15
miles
71°00′

Figure 1

Quaternary Stratigraphic Column, Southeastern Quebec

Age	Unit	Chronologic Control
Late Wisconsin	post-Lennoxville sediments	12,640 ± 190 (GSC-312; peat) 12,570 ± 220 (GSC-419; peat) 11,500 ± 160 (GSC-475-2; marine shells) 9,180 ± 180 (GSC-856; wood in colluvium) 12,000 ± 230 (GSC-936; marine shells) 11,200 ± 200 (GSC-1248; peat) 11,000 ± 240 (GSC-1289; peat) 11,200 ± 160 (GSC-1294; peat) 14,900 ± 220 (GSC-1339; peat)
	Lennoxville Till	
Middle Wisconsin	Gayhurst Formation	>20,000 B.P. (GSC-1137) ca. 4000 varves
	Chaudière Till	
Early Wisconsin	Massawippi Formation	>54,000 B.P. (Y-1683) >41,500 B.P. (GSC-507) >40,000 B.P. (GSC-1084)
	Johnville Till	
Sangamon and/or Illinoian	pre-Johnville sediments	

Stop Descriptions

Stop #1: Grande Coulée Section

Lennoxville Till; 5 m - 7 m; At east end of the section till is sandy and silty and oxidized to its base; sandy facies pinches out on the west side of the section and is represented by a compact, grey, unoxidized till facies typical of most exposures of Lennoxville Till in the region; sandy facies may represent an end moraine deposited in the proglacial lake which abutted the Lennoxville glacier during its retreat down the Chaudière valley; fabrics indicate ice flow from N to N 40°W.

Gayhurst Formation (?); 7 m thick; Interlaminated sand, silt, and clay, tentatively correlated with Gayhurst Formation sediments occurring farther south in the Chaudière valley, on the basis of stratigraphic position beneath Lennoxville Till; Chaudière Till, which underlies Gayhurst Formation sediments elsewhere, has not been identified in this section; note alternating oxidized and unoxidized beds throughout the section.

Massawippi Formation; 0.4 m thick; Fissile, structureless sand contains finely divided plant fragments dated >40,000 B.P (GSC-1084); Massawippi Formation contains pollen (predominantly *Picea* and *Pinus*) which suggest presence of a northern boreal forest at its time of deposition; i.e. a climate cooler than present; Massawippi Formation is exposed only in a lens at the west end of the section where it rests on a diamicton about 0.3 m thick. This diamicton can be traced to the east end of the section at about the same altitude and is used as a marker bed to clarify the relationships of older and younger stratigraphic units to the Massawippi Formation.

Johnville Till; 1.6 m thick; Compact, non-calcareous, grey till with strong fabric from N 40° W; only occurs in the central part of the section where it is directly overlain by the diamicton which underlies the Massawippi Formation; Johnville Till is thought to be of early Wisconsin age and correlative with the Bécancour Till of the St. Lawrence Lowlands.

Pre-Johnville Sediments;

A) Fluvial Gravel; 3 m thick; Coarse, massive gravel with strong stone imbrication indicating eastward current; gravel clasts are thickly coated with iron oxide at the west end of the section; unit can be traced from the west to east end of the section; overlain by the diamicton and Massawippi Formation at the west end of the section, directly by Johnville Till in the center, and by the diamicton and Gayhurst Formation (?) at the east.

B) Laminated silt and clay; 1.6 m thick; Possible glacial-lake sediment; non-calcareous throughout and oxidized to within 30 cm of its base; traceable with the same thickness and altitude throughout the section.

C) Fluvial Gravel; 1.3 m thick; Similar to (A) but clasts are not coated with iron oxide and are finer; rare erratics of gneiss from the Canadian Shield (70 mi to the north) found in this gravel, indicating derivation from a glaciated terrain.

D) Coarse sand grading downward into fine sand with clay partings; >4.4 m; May be a lacustrine or flood-plain sediment; unoxidized fine sand below river level contains finely-divided plant debris.

The pre-Johnville sediments are the oldest sediments recognized in southeastern Quebec and are tentatively assigned a pre-Wisconsin age. They may be the most complete record yet described of pre-Bécancour-Johnville deposition east of Toronto.

Stop #2: Chaudière River Section north of St. Martin-de-Beauce

Lennoxville Till; 18 m thick; Compact, grey, fissile, calcareous, pebbly, silty till; oxidized upward 5 m from base and downward 3 m from top; selenite crystals, apparently formed during breakdown of pyrite and calcite, were found near the base of the unit; fabrics indicate that the Lennoxville glacier was flowing from N 60° W - N 80° W during deposition of entire unit.

Lennoxville Till, sandy facies; 4.5 m thick; Sandy, loose, gravelly, oxidized till; similar to sandy tills of New England; texture is probably partially a result of incorporation of underlying sandy lake sediments; bottom contact irregular due to thrust faults which have carried lake sediment into the till.

Gayhurst Formation (?); 7 m thick; Interlaminated fine sand and clay; upper portion of the sediment is disturbed and sheared into overlying sandy till.

Chaudière Till; > 1m thick; Grey, hard, calcareous, sparingly stony till with shear structures; fabric suggests that depositing glacier moved from about N 20° E.

Stop #3: Samson River Section

The area traversed to reach the Samson River section is on the axis of the indicator ribbon of ultrabasic rocks which extends southeast from the Thetford Mines area. This indicator train is very narrow and does not have the classic "fan" shape of indicator trains described in New England. Boulder piles on the field at the start of the traverse have 10% and 14% ultrabasic cobbles, but ultrabasic frequency drops to less than 2% 6 miles north and south of the field (the width of the source area outcrop is 14 miles).

The petrology of the upper tills at the Samson River section has been studied in detail. 155 till samples from a 10 m x 4 m vertical face have had complete textural analyses (on particles <6 mm diameter); clay-mineral analyses, calcite/dolomite (Chittick) analyses, and <2µ carbonate (X-ray) analyses. The weight percents of magnetite and total heavy minerals in the fine-sand fractions, and trace element concentrations (Ni, Cr, Zr, Ti, Cu, V) in the <63µ fraction have been calculated for most samples. The purpose of this exercise was to evaluate the vertical and lateral variations of these parameters in relation to primary (sedimentation) and secondary (weathering) processes. Preliminary results of this study will be discussed at the section.

- Stratigraphy: -

Lennoxville Till

A. Till; 1 m to 3 m thick; Sandy, compact, pebbly till, oxidized to base; includes lenses of lake sediment sheared up from beds immediately at base; fabric and high ultrabasic pebble frequencies indicate that this unit was deposited by a glacier flowing from N 50° W - N 80° W.

B. Lake Sediment; 0 m to 1. 6 m thick; Highly contorted and sheared, interlaminated fine sand and clay; oxidized where less than 0.5 m thick.

C. Till; 5 m thick; Compact, grey, pebbly, silty till with a boulder pavement at the top; fabrics and pebble lithologies indicate the unit was deposited by a glacier with flow from N 20° W - N 30° W; oxidation extends along joints almost to the base; underlying deformed, clayey lake sediments grade gradually upward into till.

Gayhurst Formation; 13 m thick; Grey, calcareous, interlaminated silt and clay; contains several 0.3 m- to 1.0 m-thick zones of massive silt-clay or stony till-like sediment; the stony bands contain convolute laminations, and thin, horizontally bedded silt and clay laminae above and below them are undisturbed; the massive zones are interpreted as "turbidity" current deposits resulting from slumping from basin sides or an ice front or subaqueous slumping of sediments already deposited on the lake floor.

Chaudière Till; 1.5 m thick; Grey, calcareous, compact, clayey till; although the fabric is similar to those in the higher Lennoxville Tills (NW) the unit contains no ultrabasic pebbles; clayey texture caused by incorporation of clayey lake sediments which crop out at the base of the section; till includes clasts of these sediments and has a maroon tinge which is typical of the lake sediments.

Pre-Chaudière Sediments

A. Fine Gravel; 3 m thick; Fine gravel with large-scale crossbedding dipping south (opposite present river flow); gravel contains 5 cm to 10 cm rounded clasts of underlying lacustrine laminated silt and clay.

B. Lake Sediment; >1 m thick; Interlaminated, calcareous, silt and clay; clay laminae are chocolate brown to maroon at this locality; this unit crops out at several places just above and below river level along the Samson River.

Approximately 0.25 miles upstream from the section a diapir of the lowest (maroon) silt and clay can be seen intruding the overlying fine gravel. The origin of this structure is not known, but it illustrates the high plasticity of clay and silt in the Lac-Mégantic area.

Stop #4: Gayhurst Dam Borrow Pit

The section described here is a composite of the borrow pit, the deep gully on the far (east) side of the river, and a borehole drilled to bedrock just north of the gully. The stratigraphy in the borehole was determined by taking split-tube and Shelby tube samples at 1 m intervals and at changes in lithology. The three sites serve as the type section of the Gayhurst Formation.

- Stratigraphy: -

Lennoxville Till; ~10 m thick; Clay-till member of Lennoxville Till; compact, grey, calcareous clayey till with less than 10% sand and coarser fragments; note that there are no boulders in the till and that the till pebbles include no granodiorite component; the ground surface behind the section is mantled with large, predominantly granodiorite boulders - a one-boulder-thick ablation deposit; Fabric is N 20° E.

Gayhurst Formation

A. Upper Member; 6.2 m thick; Calcareous, laminated silt and clay; bottom 0.3 m is contorted with ball and pillow structures and convolute lamination suggesting subaqueous slumping; about 600 graded couplets are present in this unit.

B. Middle Member; 22 m thick; Fine sand grading upward to coarse sand and fine gravel with large-scale crossbedding; the top of this unit (altitude = 369 m a.s.l.) is thought to approximate the altitude of the water surface of the low level phase of Glacial Lake Gayhurst at this site; the upper contact can be traced at approximately 370 m several miles downstream from here.

C. Lower Member; 59 m thick; Fine-grained lacustrine sediment; contains approximatley 3400 graded, calcareous silt and clay couplets; upper 7 m and lower 10 m are very fine sand and silt; graded couplets contain sparse, finely disseminated plant fragments; several zones show evidence of subaqueous slumping.

Chaudière Till; 11.8 m thick; Two grey, compact, calcareous, silty, sandy till units separated by 2.4 m of highly deformed, brecciated silt and clay laminae; upper part of till derived from northwest; lower part derived from north or northeast.

Pre-Chaudière Sediments

A) Lake sediment; 2.4 m thick; intensely deformed laminated silt and clay.

B) Fine Gravel; 1.2 m thick; fine gravel with abundant clasts derived from sources to northwest.

C) Johnville (?) Till; 2.7 m thick; stony, compact sediment; no sample recovered.

D), Lake sediment (?); 1.8 m thick; silt and clay; no sample recovered.

E) Weathered bedrock (?); 1.8 m thick; no sample recovered.

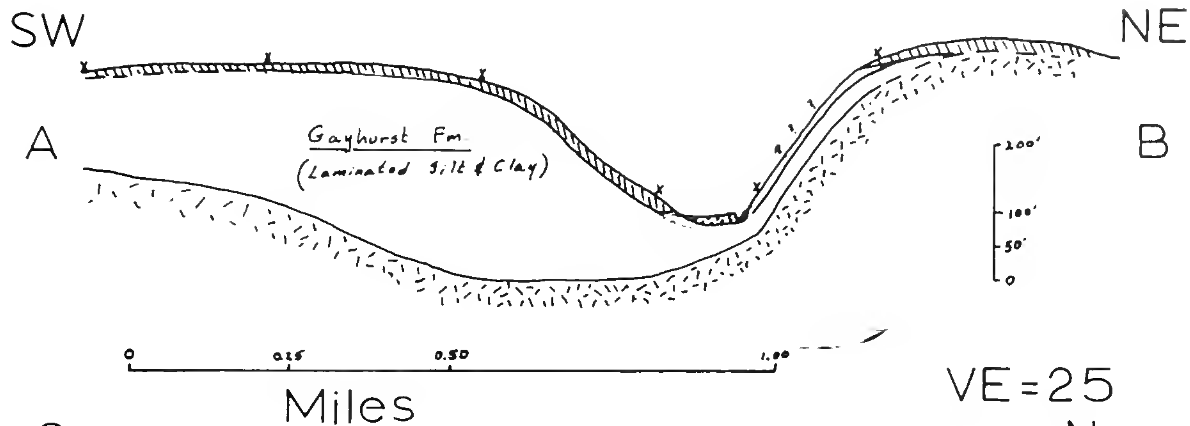
F) Pyritiferous black slate (Compton Fm., Devonian (?); 2.8 m of core recovered.

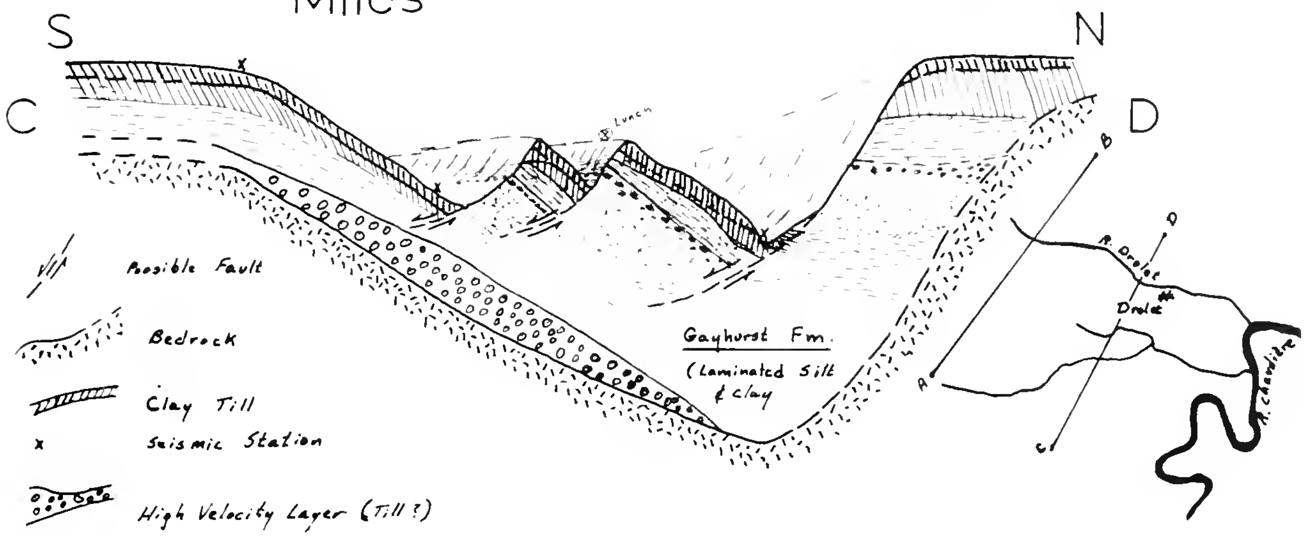
Stop #5: Drolet Slump Blocks

The asymmetric ridges on which we will eat lunch are thought to be large, coherent slump blocks composed of Gayhurst Formation sediments capped by the clayey lentil of Lennoxville Till. The blocks have dropped as much as 50 m vertically along normal fault planes, appparently in response to erosion by the Chaudière River. Although rotational slumping is quite common everywhere in the Lac-Mégantic region, these blocks are unusual for their size (more than 1 mile long) and location more than two miles from the site of river erosion.

The Drolet River occupies a broad, deep bedrock depression (see accompanying figures).

Rivière Drolet, P.Q. Cross Sections





Diagrammatic Cross Section of Rotational Slump: Rivière Chaudière, P.Q.

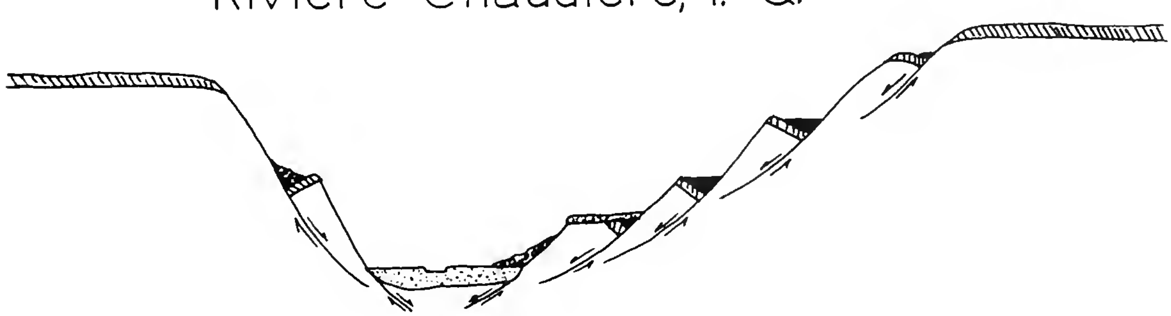


Figure 2

Stop #6: Mégantic Moraine Complex

The smooth ridges at this locality are thought to be till moraines deposited during a halt of the Lennoxville glacier. The till exposed in the cut through the moraine ridge is quite sandy and loose and the complex grades into ice contact stratified deposits at lower altitudes, several tens of meters to the northwest.

Smooth till or gravel ridges (usually nested) are common moraine forms in the Lac-Mégantic region, but they are generally less than a mile long. It is not known whether the present ridge form is entirely a product of ice deposition or whether the form was accentuated by meltwater flowing between ridges after their formation.

Ice-front positions during retreat of the Lennoxville glacier have been drawn by connecting linear trends of assemblages of ice-front features such as till moraine ridges, ice-contact subaerial outwash, ice-contact deltaic sediments, and meltwater channels.

Stop #7: Striae West of Lac-Mégantic

The striae on this outcrop of interbedded Compton Formation slate and sandstone reflect two directions of movement for the Lennoxville glacier. The "tails" of rock in the lee of pyrite cubes indicate the sense of movement - toward the Boundary Mountains (SE).

Outcrops in the Lac-Mégantic region commonly show two or more distinct azimuth groupings of striae with very few intervening directions.

Note the "offset" striae on this outcrop; one part of the outcrop has been uplifted a couple of centimeters above the other after the striae were cut. Such offsetting is common in this region. Failure is along bedding planes in the slate. On the basis of such evidence, Oliver, Johnson and Dorman (1970) have suggested that the cumulative postglacial displacement along such faults may have been "substantial".

Stop #8: Woburn Esker Gravel Pit

This pit has exposed in the past an excellent assemblage of structural features resulting from melting of buried ice. The pit face exposes two main areas of slumping over melting ice. Note that the faults formed in response to subsidence are high-angle reverse faults with small compensating normal faults. The significance of this observation will be discussed briefly.

This stop is about 1½ to 2 miles from the Coburn Gore customs houses. The field trip ends here and those travelling to Rangely-Stratton need only follow route 34 east to the border where it joins the Arnold Trail (rte. 27) which passes through Rangely and Stratton. Drive very carefully as this road is tortuous, narrow, and heavily travelled by logging trucks.

References

Behling, R.E., 1965, A detailed study of the Wisconsin stratigraphic section of the upper Lamoille vally, north central Vermont: unpub. M.S. thesis, Miami University, Oxford, Ohio, 125 p.

Borns, H.W., Jr., 1966, An end-moraine complex in southeastern Maine: Geol. Soc. America, Northeastern Sec. Ann. Mtg., Philadelphia, p. 13-14.

Borns, H.W., 1963, Preliminary report on the age and distribution of the late Pleistocene ice in north central Maine: Amer. J. Sci., v. 261, pp. 738-740.

Brady, J.G. and Dean, R.S., 1966, The composition and properties of ceramic clays and shales of Québec; Dept. of Energy, Mines and Resources, Mines Branch, Ottawa, Research Report, R 187, 107 p.

Cannon, William F., 1964, The petrography of the tills of northern Vermont: unpub. M.S. Thesis, Miami University, Oxford, Ohio.

Chalmers, R., 1898, Surface geology and auriferous deposits of southeastern Quebec: Geol. Surv. of Canada, Ann. Report, vol. 10, pt. V., 160 p.

Cooke, H.C., 1937, Further note on northward moving ice: Am. Jour. Sci., 5th ser., v. 34, p. 221.

Duquette, G., 1960, Preliminary report on Gould area: Quebec Dept. Mines, Prelim. Rept. 432, 10 p.

Flint, R.F., 1951, Highland centers of former glacial outflow in northeastern North America: Geol. Soc. America Bull., v. 62, p. 21-38.

Gadd, N.R., 1960, Surficial geology of the Bécancour map-area, Quebec: Geol. Surv. Canada, Paper 59-8, 34 p.

Gadd, N.R., 1964, Surficial geology, Beauceville Map-area, Quebec: Geol. Surv. Canada, paper 64-12.

Gadd, N.R., 1964, Moraines in the Appalachian region of Quebec: Bull. Geol. Soc. Amer., vol. 75, pp. 1249-1254.

Gadd, N.R., 1965, Surficial geology, Chaudière River valley: Geol. Survey of Canada, Paper 65-1, pp. 115-117.

Gadd, N.R., 1965, Buried valleys and gold placers, Beauce County: Geol. Surv. Canada, Paper 65-1, pp. 117-119.

Gadd, N.R., 1967, The Pattern of Glacial Recession in Southeastern Québec: (Abs.), Northeast section of the Geol. Soc. of America 1967, Annual meeting, pp. 28-29.

Gadd, N.R., (in press), Pleistocene geology of the Central St. Lawrence Lowland: Geol. Surv. Canada, Memoir 359.

Gorman, W.A., 1955, Preliminary report on St. Georges-St. Zacharie area: Quebec Dept. Mines, Prelim. Rept. 314, 5 p.

Lasalle, P., 1966, Late Quaternary vegetation and glacial history in the St. Lawrence Lowlands, Canada: Leidse Geol. Mededel., pt. 38, p. 91-128.

McDonald, B.C., 1966, Pleistocene geology studies, Richmond-Sherbrooke region, southeastern Québec: Geol. Surv. of Canada, Paper 66-1, p. 167.

McDonald, B.C., 1966, Surficial geology, Richmond-Dudswell, Quebec: Geol. Surv. Canada, Map 4-1966.

McDonald, B.C., 1966, Auriferous till in the Eastern Townships, southeastern Quebec: Geol. Surv. Canada, Paper 66-2, p. 51-54.

McDonald, B.C., 1967, Pleistocene events and chronology in the Appalachian region of southeastern Quebec, Canada: Ph.D. dissert., Yale Univ., New Haven, 161 p.

McDonald, B.C. 1967, Surficial geology, Sherbrooke-Orford-Memphremagog, Quebec: Geol. Surv. Canada, Map 5-1966.

McDonald, B.C. 1967, Wisconsin stratigraphy and ice-movement directions in southeastern Quebec, Canada, (abs.): Geol. Soc. America, Northeastern Sec. Ann. Mtg., Boston, p. 41-42.

McDonald, B.C., 1967, Glacial geology of Eaton River watershed and adjacent areas of southeastern Quebec: Geol. Surv. of Canada Paper 67-1, pp. 166-167.

McDonald, B.C., 1968, Deglaciation and differential postglacial rebound in the Appalachian region of southeastern Quebec: Jour. Geology, V. 76, p. 664-677.

McDonald, B.C. 1968, Composition of some garnets used in provenance studies of tills; Jour. Sed. Pet., vol. 38, no. 3, pp. 956-957.

McDonald, B.C. 1969, Surficial geology of La Patrie-Sherbrooke area, Quebec, including Eaton River watershed: Geol. Surv. Canada, Paper 67-52, 21 p.

McDonald, B.C. and Shilts, W.W., (in press) Quaternary events and stratigraphy, southeastern Quebec: Geol. Soc. America Bull.

McGerrigle, H.W., 1936; Gold placer deposits of the Eastern Townships: Que. Bur. Mines, Annual Rept. for 1935, pt. E.

Marleau, R.A., 1968, Woburn-East Megantic-Armstrong area, Frontenac and Beauce counties: Quebec Dept. Nat. Res., Geol. Rept. No. 131, 55 p.

Oliver, J., Johnson, T., and Dorman, J., 1970, Postglacial faulting and seismicity in New York and Quebec: Can. Jour. Earth Sci., V. 7, no. 2, pt. 2, p. 579-590.

Prest, V.K., Grant, D.R., and Rampton, V.N., 1967, Glacial Map of Canada, Geol. Surv. of Canada, Map 1253A.

Shilts, W.W., 1967, X-Ray diffraction as a stratigraphic tool in the Pleistocene of Northern Vermont and southeastern Québec: (Abs.) Program, Geol. Assn. of Canada Internationa Meetings, Kingston, Ontario, 1967.

Shilts, W.W., 1968, Pleistocene geology, Lake Mégantic area, southeastern Québec: Geol. Surv. of Canada, Paper 68-1, p. 123.

Shilts, W.W., 1969, Quaternary geology of the upper Chaudière River drainage basin, Quebec: Geol. Surv. Canada, Paper 69-1A, p. 218-220.

Shilts, W.W., 1969, Pleistocene geology of the Lac Mégantic region, southeastern Quebec, Canada: Ph. D. dissert., Syracuse Univ., Syracuse, 154 p.

Shilts, W.W., 1970, Indicator studies; Pleistocene geology; Lac Megantic region, Quebec: Geol. Surv. Canada, Paper 70-1A, p. 185-186.

Shilts, W.W., (in press) Geochemical anomalies in till; southeastern Quebec: Geol. Surv. Canada, Paper 70-1B,

Stewart, David, P., 1961, The glacial geology of Vermont: Vermont Geol. Survey, Bull. No. 19, 124 p.

Stewart, D.P. and MacClintock, P., 1964, The Wisconsin stratigraphy of northern Vermont: Am. Jour. Sci., v. 262, p. 1089-1097.

Stewart, D.P., and MacClintock, P., 1969, The surficial geology and Pleistocene history of Vermont: Vermont Geol. Surv., Bull. 31, 251 p.

Terasmae, J., 1958, Non-glacial deposits in the St. Lawrence Lowlands, Quebec; Geol. Surv. of Canada, Bull. 46, contributions to Canadian Palynology, Part II, pp. 13-28.

Terasmae, J., 1965, Geological Survey palynological studies: Geol. Surv. Canada, Paper 65-1, p. 158-159.

Thornes, J.B., 1965, Some observations on the late-glacial stages in the Coaticook valley, southern Quebec: Cahiers de Géog. de Québec, 9th year, no. 18, p. 223-238.

Van Ingen, R., 1967, (Abs.) Granite Studies related to ore formation in the Eastern Townships, Québec; in Proceedings, Symposium on Geochemical Prospecting, Ottawa, April, 1966; Geol. Surv. of Canada, paper 66-54, pp. 277.278.

Quaternary History of Northwestern Maine *

Harold W. Borns, Jr., University of Maine
Parker E. Calkin, State University of N. Y. at Buffalo

The highlands of northwestern Maine, including the Longfellow and Boundary Mountains, were overridden at least twice and perhaps several more times, by continental ice sheets during the Quaternary Period. These episodes are indicated by five, widely separated exposures displaying two-drifts sequences composed of lodgment tills separated by lacustrine and fluvial sediments. The freshness of these drifts suggest that they are probably of Wisconsin age, however this is equivocal as no way has been found to assign absolute ages to them.

Caldwell (1959) reports a two-till sequence at New Sharon, in central Maine, separated by organic materials dated at more than 38,000 years old. A recent C^{14} age determination (Stuiver, personal communication), and an analysis of the wood fragment which shows that the tree was crushed while still green, indicate that the organic material at New Sharon was overridden by ice more than 52,000 years ago. Presently there is no way of determining the relationship of the New Sharon sequence and the undated sequences to the northwest.

The last ice sheet, whose retreating margin stood along the present Maine coast approximately 13,500 years ago (Borns, unpublished), thinned, separated and stagnated over the Longfellow and boundary Mountains of northwestern Maine, a belt 60 miles wide and rising over 3000 ft. above bordering lowlands. Nearly contemporaneous stagnation, throughout and perhaps to the southeast of the mountains, is evidenced by the distribution and volume of ice-contact stratified drift. Coupled with this is the lack of evidence of a receding active ice margin.

The separation of this ice in Maine from the still-active receding ice sheet immediately to the northwest in Quebec, occurred approximately 12,800 years ago and the subsequent dissipation of stagnant ice in the mountains was complete by approximately 12,000 years ago.

The highest glacial cirques in northwestern Maine on Crocker Mtn., with floors at an altitude of approximately 2700 ft., reveal no evidence of glacial reactivation during and subsequent to the dissipation of the last ice sheet.

---

* This summary is the result of current research being carried on under National Science Foundation Grant Ga-1563 to the University of Maine. The geological statements in the Road Log are brief and subject to change because the research is still in progress.

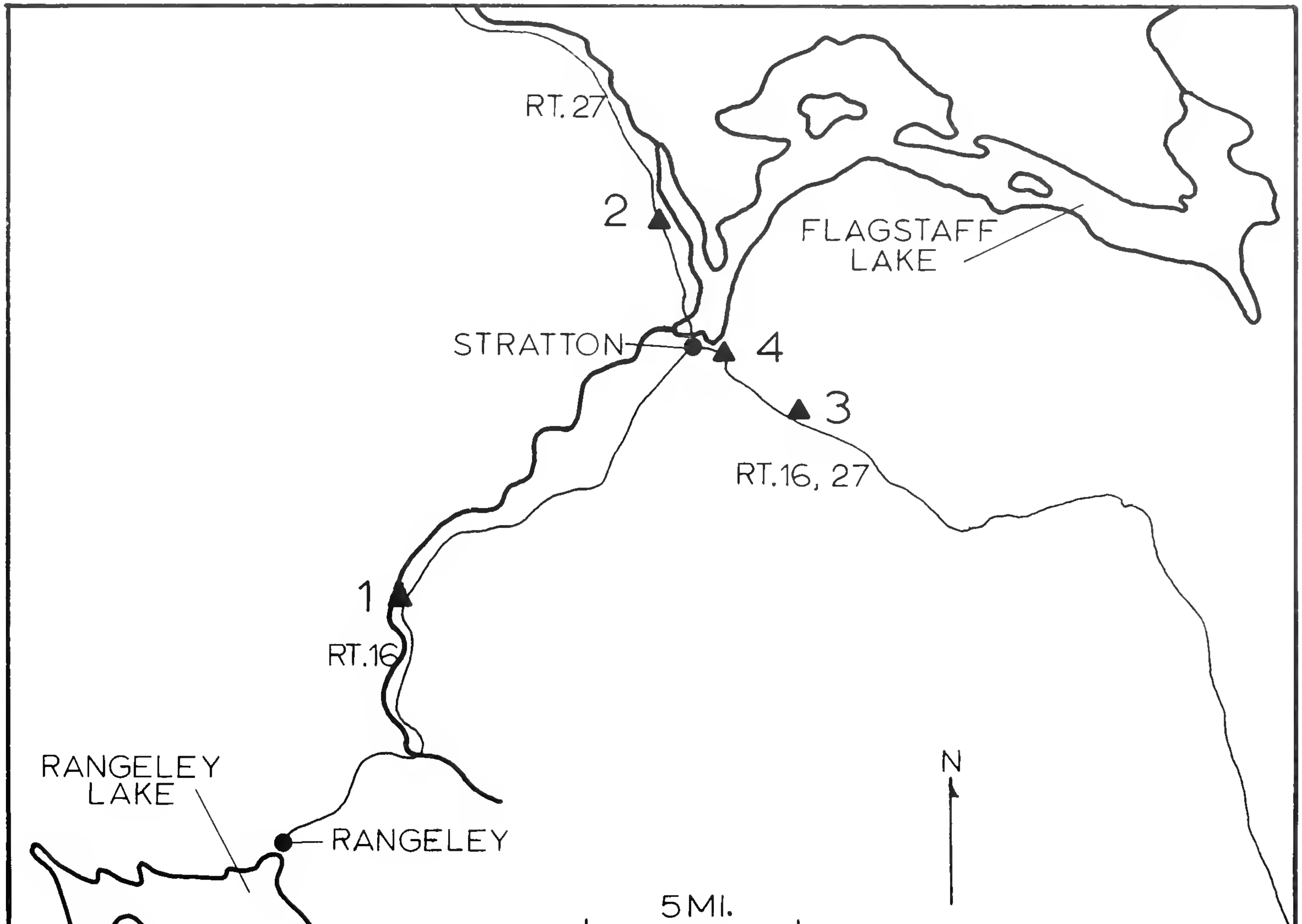


Fig. 1. Index map showing routes and stop locations.

Topographic 15 minute Quadrangle Maps:

Kennebago Lake
Rangeley
Stratton

Purpose of the trip: To examine some evidence of (1) multiple glaciation and (2) the mode of dissipation of the last ice sheet in northwestern Maine.

Road log. Mileage from Rangeley Center.

Time 7:45 A.M., Sunday, October 4.

0 miles	Leave Rangeley Center. Travel toward Stratton on Rt. 16.
4.5	Travel on crest of esker for 1 mile.
5.5	Descend from Esker
7.5 (1 hour)	STOP 1. The 50 foot high cut bank on the east bank of the South Branch of the Dead River exposes a stratigraphic section composed of till at the base overlain successively by glaciolacustrine silts and sands, fluvial gravels and a second till. This sequence represents two glacial episodes separated by a nonglacial episode.
17.5	Stratton; travel north towards Eustis on Rt. 27.
21.0	Turn left into Cathedral Pines Camp Ground and stop at public beach.
21.5 (45 min.)	STOP 2. A discussion of the lake history of the "Flagstaff Lake - South Branch" basin. Flagstaff Lake occupies a part of the basin that formerly held the slightly larger Glacial Lake Bigelow (Leavett and Perkins, 1935). Recent study has revealed a complex glacial lake history for this basin which not only contains Flagstaff Lake, but which extends southwestward approximately 10 miles, to STOP 1. The South Branch of the Dead River presently flows along the axis of this section of the basin. As deglaciation progressed, large ice masses were left stagnating in this as well as in many other basins of the region. Evidence, primarily in the form of deltas, for former lake levels up to 400' above the maximum level of Flagstaff Lake is present in the basin. These lake levels were controlled by ice blockage and by the successive uncovering of lower spillways as the stagnant ice mass or masses dissipated. Lake water was derived from meltwater draining from adjacent basins primarily by way of the Kennebago Lake basin and the North Branch of the Dead River as well as from melting of the ice within the basin.

Glacial Lake Bigelow , the last glacial lake known to occupy the basin, filled it to a present altitude of approximately 1200 ft., roughly 40 ft. above the present maximum level of man-made Flagstaff Lake. This glacial lake was completely drained into the Kennebec River via the Dead River when it overflowed and cut away a threshold of till at the site of Long Falls Dam on the northeast end of Flagstaff Lake. This event occurred some time after approximately 12,800 years ago (Borns and Hagar, 1965; Borns and Stuiver, unpublished).

Cathedral Pines Camp Ground rests upon a delta built into Glacial Lake Bigelow by meltwater from the North Branch of the Dead River. At that time ice masses still occupied the valley of the North Branch as indicated by the numerous kettle holes present in the upstream equivalent of the delta sands. Subsequently Glacial Lake Bigelow was drained and the delta was dissected by both the North and South Branches of the Dead River.

22.0 Rt. 27: travel south to Stratton on the dissected Glacial Lake Bigelow delta.

25.5 Stratton: travel south on Rt. 16 and 27 on delta surface for 1 mile and parallel to an esker for 2 miles.

28.5 (30 min.) STOP 3. A pit in the "Stratton Esker". A discussion of the significance of eskers in this region.

Several recent borrow pits in this esker have exposed excellent examples of grain-size distribution, stratification and ice-contact collapse structures typical of "classic eskers".

Considering the internal structures, the segmentation, the relationship of the esker to the regional topography and the provenance of the sediments it is concluded that this esker was formed at the base of stagnant ice, probably during the late nunatak stage, and that meltwater was channeled in the ice along the present course of the North Branch of the Dead River, across the Flagstaff Lake basin and along the trend of the Stratton esker into the Carrabassett basin.

It can be demonstrated that in this region esker channels carried meltwater from ice mass to ice mass across nunataks proving that valley-filling ice masses existed contemporaneously throughout the region. Therefore the mode of dissipation of the last ice sheet was thinning, separation and stagnation on a broad scale within this mountainous belt.

28.5 Leave STOP 3. Travel towards Stratton on Rt. 16 and 27.

30.8 Cross stream.

31.0 Right on 1st road past stream.

31.2 STOP 4. A pit in the "Stratton Esker".
(25 min.)

At this location, as well as at several others nearby, deltaic sand overlies the "Stratton Esker". This evidence coupled with the presence of a few kettle holes in the delta and in the fluvial sands in the North Branch of the Dead River which are the upstream equivalent of the delta sands, suggests that at the time of Glacial Lake Bigelow residual ice was still present in very minor amounts. The major part of regional deglaciation has been accomplished.

31.2 Return to Rangeley via Stratton on Rt. 16.

49.7 Rangeley

References Cited

Borns, H. W. Jr., and Hagar, D. C., 1965, Late-glacial stratigraphy of a northern part of the Kennebec River Valley, western Maine: Geol. Soc. American Bull., v. 76, p. 1233-1250.

Caldwell, D. W., 1959, Glacial lake and glacial marine clays of the Farmington Area, Maine-origin and possible use as lightweight aggregate: Maine Geol. Survey, Special Geol. Studies Ser., No. 3, 48 p.

Leavitt, H. W., and Perkins, E. H., 1935, Glacial geology of Maine, v. 2: Maine Technology Expt. Sta. Bull. 30, 232 p.

SULFIDE MINERALIZATION IN PRE-SILURIAN ROCKS, THRASHER PEAKS, WESTERN MAINE

René Fournier
Department of Geology
University of Cincinnati

The area at Thrasher Peaks affords an opportunity to observe several aspects of the geology of northwestern Maine. The present trip will include examination of the lithologies and structure in the Magalloway Member of the Dixville Formation and some of the features of the sulfide mineralization present in this member. There will also be a short traverse to study the relationship of the Magalloway Member to some younger (?), well-preserved agglomerates, arkoses, slates and Devonian microgranite.

Thrasher Peaks, a series of low, northeast trending hills, lies on the northwest limb of the Boundary Mountain anticlinorium. As indicated on geologic maps by Green (1968) and Harwood (1966, 1969), the southeastern slopes of the Peaks are underlain by the Magalloway Member of the Middle Ordovician Dixville Formation and Devonian microgranite. Green's map shows a small detached body of Seboomook slate at the southwest end of this slope. The northwestern slopes are underlain by the Devonian Seboomook Formation and the crest-line of the Peaks is underlain by Ordovician greenstones and agglomerates, Silurian conglomerates and Devonian microgranite. Assignment by Green of the greenstone and agglomerate unit to the Ordovician and the conglomerates to the Silurian is tentative. A structure section through the area by Green (1968) shows Thrasher Peaks as a minor anticline, slightly over-turned to the southeast. Details of the structure and regional stratigraphy may be found in articles by Green (1968), Green and Guidotti (1968), and Harwood (1966, 1969; also Harwood and others, this guidebook).

With the exception of a few additional rock types located during preliminary work by the author, the local stratigraphic sequence is that of Green, as summarized below.

Magalloway rocks in the area consist predominantly of quartz-feldspar-sericite schist with dark gray, thinly laminated chloritic slates and light green phyllites as minor elements. There are also lenses of dark gray siliceous metagreywacke and black chert in the dark gray slates. The dark slates commonly are highly distorted, especially at the contacts with the siliceous units described above. The quartz-feldspar-sericite schist is medium to coarse grained and at places may be aptly described as "augen schist". The protolith of the schists is problematical due to the extreme shearing in the rock; some are certainly clastic sediments whereas others, as suggested by Green (1968), have been derived from felsic volcanics.

The greenstone and agglomerate unit which crops out on the ridge-line of Thrasher Peaks contain cobbles of quartz-feldspar grit which are similar in mineralogy and texture to the quartz-feldspar schists of the Magalloway. The relationship is complicated, however, by the presence of an arkosic unit in contact with the agglomerates, which although similar to the quartz-feldspar-sericite schist of the Magalloway, is relatively

unmetamorphosed and lacks evidence of the extreme deformation found in the Magalloway rocks. As the arkosic unit is in contact with the agglomerates, this unit would appear to be a more likely source for the grits in the agglomerates than the quartz-feldspar-schist of the Magalloway.

In addition to the presence of the arkosic grits in the agglomerate, rounded to subrounded cobbles and boulders of these grits are present in a thin-bedded slate which lies between Magalloway rocks and the arkosic unit described above. Green mapped these slates as Seboomook. The arkosic cobbles and boulders are randomly and sparsely distributed throughout the slates. In addition to the boulders there are several large exotic blocks of microgranite. The source of the arkosic boulders is probably the arkosic unit mentioned above; the source of the microgranite is probably the large microgranite body which crops out about 500 feet from the microgranite blocks.

A suitable mechanism to account for the presence of the boulders of arkose and exotic microgranite in the slate has yet to be worked out. Movement of the slate unit through gravity sliding, submarine slumping or mudflow are possibilities. There also is the possibility that the exotic microgranite blocks represent slices of the nearby microgranite body taken during thrusting. The exact nature of the mechanism rests on the age relationship of the slate to the arkose unit, to the microgranite and to the Magalloway rocks.

Sulfide mineralization in the Thrasher Peaks area is present in the metasediments and metavolcanics of the Magalloway Member. Presence of copper, iron, lead and zinc sulfides is of interest as they appear to be genetically related to volcanic activity during the period of deposition of the Magalloway. Post depositional deformation mobilized the sulfides into fold hinges where they are presently concentrated.

The area of mineralization may be defined approximately by a narrow linear zone which parallels the regional structure along the southeastern flanks of Thrasher Peaks. Known limits of this zone are about 300 feet in width and more than three miles in length. Significant quantities of sulfides, however, are known at only a few locations within the zone. All sulfide mineralization appears to be limited to the quartz-feldspar sericite schists and chloritic slates of the Magalloway Member. Minor pyrite and chalcopyrite is associated with the greenstones and agglomerate unit described previously.

Mineralization is present either as massive accumulations associated with fold hinges or disseminated in the host rocks. Disseminated sulfide is predominently pyrite with very minor chalcopyrite. Commonly the pyrite in the schist is small (1-5 mm),

angular to subangular grains arranged parallel to the foliation of the rock. Pyrite is also present as fine grained masses, 1/4" to 1/2" across, and in cross-cutting quartz stringers. In the slates, fine grained aggregates of pyrite are present parallel with the bedding of the slate. These masses, generally 1/2" to 3" across, are lenticular in shape. Finer, thin stringers of pyrite are parallel with the cleavage.

The larger sulfide masses are simple in mineral composition: pyrite, chalcopyrite and sphalerite. Galena generally is not associated with these sulfides, but is found in discrete thin layers close to these sulfide masses. Pyrite is ubiquitous, present as fractured cubes and subangular grains. These grains exhibit a wide range in size, from microscopic fragments to grains about one centimeter in size. The fractured pyrite has been invaded by chalcopyrite and sphalerite, which are present as intermixed poorly defined bands parallel with the foliation of the host rock. Rare chalcopyrite inclusions are found in the sphalerite.

At the present stage of the study of this area little may be said of the structural aspects of the mineralization. Detailed mapping at known mineralized areas indicates, however, that the sulfides are localized in the hinges of folds. In these areas masses of sulfides have replaced the quartz feldspar sericite schist resulting in zones 20 to 40 feet across. Although the precise relationships are not yet clear, the sulfides appear to have participated in the second period of folding which has resulted in tight, southwest plunging folds. These folds are asymmetric, overturned to the southeast and with small cross section and closure.

Spatial and temporal relations of the sulfides to the host rocks suggest that deposition of the sulfides was pene-contemporaneous with the pyroclastics and clastics of the Magalloway Member. Source of the mineralizing fluids may well have been subaqueous hydrothermal venting associated with the acid volcanic activity. The fluids may have impregnated the soft, water saturated sediments with sulfides resulting in broad dissemination of the base metals. Subsequent deformation mobilized the sulfides into the hinges of folds as the folds formed.

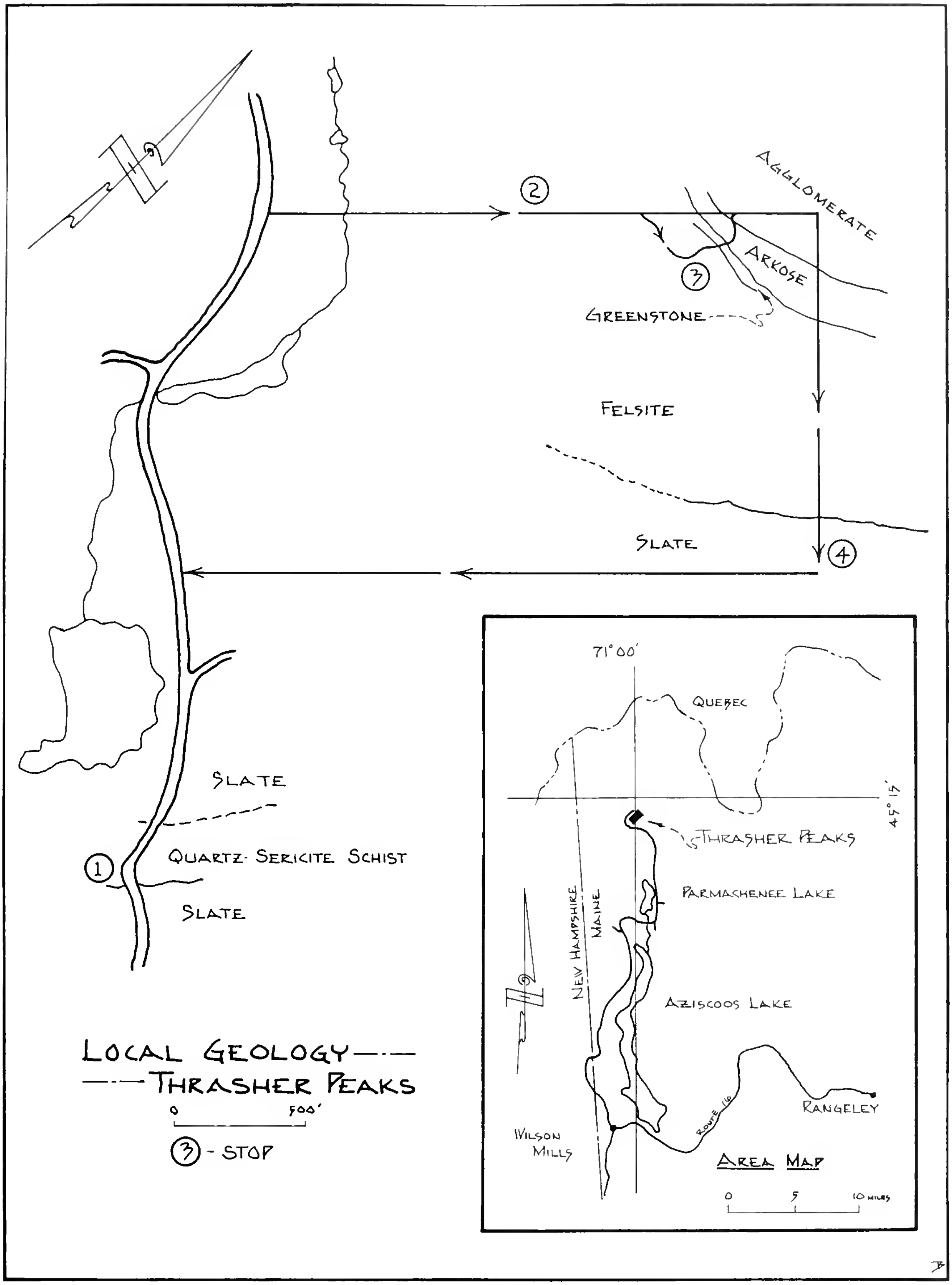
References

Green, J. C., 1968, Geology of the Connecticut-Parmachenee area, New Hampshire and Maine: Geol. Soc. America Bull., v. 79, p. 1601-1638.

Green, J. C. and Guidotti, C. V., 1968, The Boundary Mountain anticlinorium in northern New Hampshire and northwestern Maine: in Studies in Appalachian geology, Northern and Maritime, E-an Zen, W. S. White, J. B. Hadley, and J. B. Thompson, Jr., editors, Interscience, New York, p. 255-266.

Harwood, D. S., 1966, Geology of the Cupsuptic quadrangle, Maine: U. S. Geol. Survey Open File Report 850, 259 p.

Harwood, D. S., 1969, The Second Lake anticline - a major structure on the northwest limb of the Boundary Mountain anticlinorium in northern New Hampshire, west-central Maine and adjacent Quebec: U. S. Geol. Survey Prof. Paper 650-D, p. 106-115.


AGGLOMERATE
ARKOSE
GREENSTONE
FELSITE
SLATE
SLATE
QUARTZ-SERICITE SCHIST
SLATE
LOCAL GEOLOGY — THRASHER PEAKS
0 500'
3 - STOP
71°00'
QUEBEC
45°15'
THRASHER PEAKS
PARMACHENEE LAKE
NEW HAMPSHIRE
MAINE
AZISCOOS LAKE
ROUTE 16
RANGELEY
WILSON MILLS
AREA MAP
0 5 10 MILES

Road Log for Trip F-2

Assemble at the Gulf Oil station in Wilson Mills, Maine at 8:30 AM. Wilson Mills may be reached by travelling west out of Rangeley on Maine Highway 16. It is about a half an hour's drive from Rangeley.

We will follow most of the route taken earlier by Harwood (see trip A-3, this guide). The trip begins at milepost 34.2 of that log and continues to milepost 55.8; begin clocking mileage for this trip at this point.

Mileage

0.0 Continue north past old farmhouse. Parmachenee Lake is to the west; the buildings on the island belong to the Parmachenee Club.

4.6 Road comes to a "Y"; follow road to the left. Cross First East Branch of the Magalloway River. It may be better to ford the stream rather than use the bridge.

6.3 STOP 1: The road turns sharply to the northeast, skirting a large outcrop to the right. Park beyond the bend and gather at the outcrop. This stop is in the Magalloway member of the Dixville Formation established by Green (1968). We will spend about an hour here looking at the various lithologies and the sulfide mineralization.

The quartz-sericite schist at the outcrop is typical of Magalloway in the local area; in areas of extreme shearing, transposition of the bedding is typical, whereas in areas of minor deformation the grit is coarse and its clastic nature is apparent. Within 100 feet on both sides of the outcrop are crops of thinly laminated dark chloritic slates which locally are highly deformed. Lenticular bodies of dense, black chert as well as dark siliceous tabular bodies of greywacke can be located in crops of slate east of the first outcrop. Primary depositional features such as convolutions and truncations may be seen here as well.

The rocks here are structurally complex; two cleavages, one essentially with the bedding and one trending northwest may be seen. The latter is a crenulation cleavage, common in the slates. Evidence from top sense observations indicates that the quartz-sericite schist is older and forms the core of an overturned fold which plunges to the southwest.

The outcrop at the stop is heavily mineralized with pyrite, chalcopyrite and sphalerite in decreasing order of abundance. Very minor amounts of galena are also present; pyrrhotite has not been noted. The zone, which is about 20 feet across and extends across the road, is parallel with the general schistosity of the outcrop.

Reassemble and continue north on the road. We will park on the road and spend the rest of the trip on foot. The trail that we shall follow is well marked, but be careful of the sappling stumps on the trail. Proceed in a northeast direction along the trail for about 800 feet.

6.8 STOP 2: The outcrop is a felsitic rock which crops out along the southeastern flank of Thrasher peaks. Green (1968) and Harwood (1966) have mapped the unit as Devonian microgranite or intrusive felsite respectively. The unit is the largest of a series of small elliptical felsic rocks extending in a narrow belt northeastward from the New Hampshire line. The rock texture is holocrystalline carrying micro-crystals of subhedral quartz, orthoclase, plagioclase and minor muscovite. Intergrowths of tourmaline and quartz are common, as well as perthite and myrmekite. The texture becomes coarser in the interior of the outcrop zone.

The felsite is massive and very dense; the only deformation is represented by well developed joints. Some shearing is present at the lower contact which we will see later. Incipient banding may be noticed, but the cause is not clear.

Continue along the trail to the northeast for about 500 feet. At this point we will break away from the trail and proceed through the woods to the southeast for about 200 feet. We will be at the base of a large southwest-facing outcrop, the base of which is felsite.

STOP 3: Starting at the base and moving upward on the northwest side of the outcrop the felsite gives way to a thin clastic zone in which small (1 mm - 1 cm) rounded to angular fragments of quartz have been cemented by silica. The texture is best seen on the weathered surface. Fresh surfaces of this rock do not reveal the texture; rather the fresh surface is gray with dark-green mottled texture suggesting that the rock has been hornfelsed.

Above the clastic zone is a thin greenstone, best seen at the top of the outcrop. The contact between these two units is difficult to locate.

A thin, persistant arkose is present about 50 feet to the east of the last outcrop. Grains of well sorted angular fragments of quartz, orthoclase, plagioclase and very minor lithic material make up this unit. The matrix has been slightly sericitised. Thin (3 cm - 5 cm) beds of silt-sized fragments separate thicker beds of arkose. These may mark large scale cross stratification of the arkose. Outcrops of this lithology in this area are typically free from the effects of deformation. Clasts of arkose similar to this are present in the agglomerate to be seen next as well as at the next stop.

About 200 feet to the north, near the trail we have been following, is the contact of the arkose with the agglomerate. Agglomerates have been mapped by Green (1968) in this area and are tentatively placed within the Dixville. They are here composed of fragments of various

volcanic lithologies, felsites and arkose. The arkose fragments are very similar to those seen just to the south. Pyrite and minor chalcopyrite are very common in the agglomerate.

Regroup at the intersection of the trails going to the north-east and southeast. Proceed to the southeast for about 1000 feet. We will be going downhill and across fairly continuous outcrops of felsite to its lower contact. The texture of the felsite is locally relatively coarse. Small patches of felsite containing arsenopyrite are associated with one of the joint sets.

STOP 4: At the lower contact of the felsite. In this area the felsite is in contact with a dark gray slate, but farther to the east it is in contact with Magalloway quartz-sericite schist. Here the contact is faulted, although the fault zone lacks quartz veining, gouge and mylonite. The slates were mapped by Green (1968) as a thin slice of Seboomook Formation; present work has extended the area of outcrop of this slate.

Proceed southeast across the small valley to a low ridge where good outcrops of folded slate may be seen. This slate is unusual because it contains sub-rounded cobbles and boulders of arkose similiar to that seen at stop three. Early folding appears to have resulted from slumping; later deformation has resulted in broader southwest plunging folds.

We will return to the vehicles along a trail to the southwest. Proceed about 2000 feet to the southwest until it comes out on the road. The vehicles are north along this road. Return to Rangeley.

TRIP F-3

Base Metal Ore Deposits of southern Hancock County, Maine

by

Robert G. Doyle
State Geologist, Maine Geological Survey
Augusta, Maine

Assemble in private cars at 6:45 AM at Rangeley Inn Parking lot on Sunday October 4, 1970 - or - in shopping center on south side of Bucksport on U.S. Route 1 at 10:15.

This trip will visit two base metal properties in southern Hancock County, the Harborside Mine of the Callahan Mining Co. and the Blackhawk Property near Blue Hill. These localities represent two possible examples of base metal sulphide deposits representative of the ore prospect environments of the southern volcanic belt of Maine.

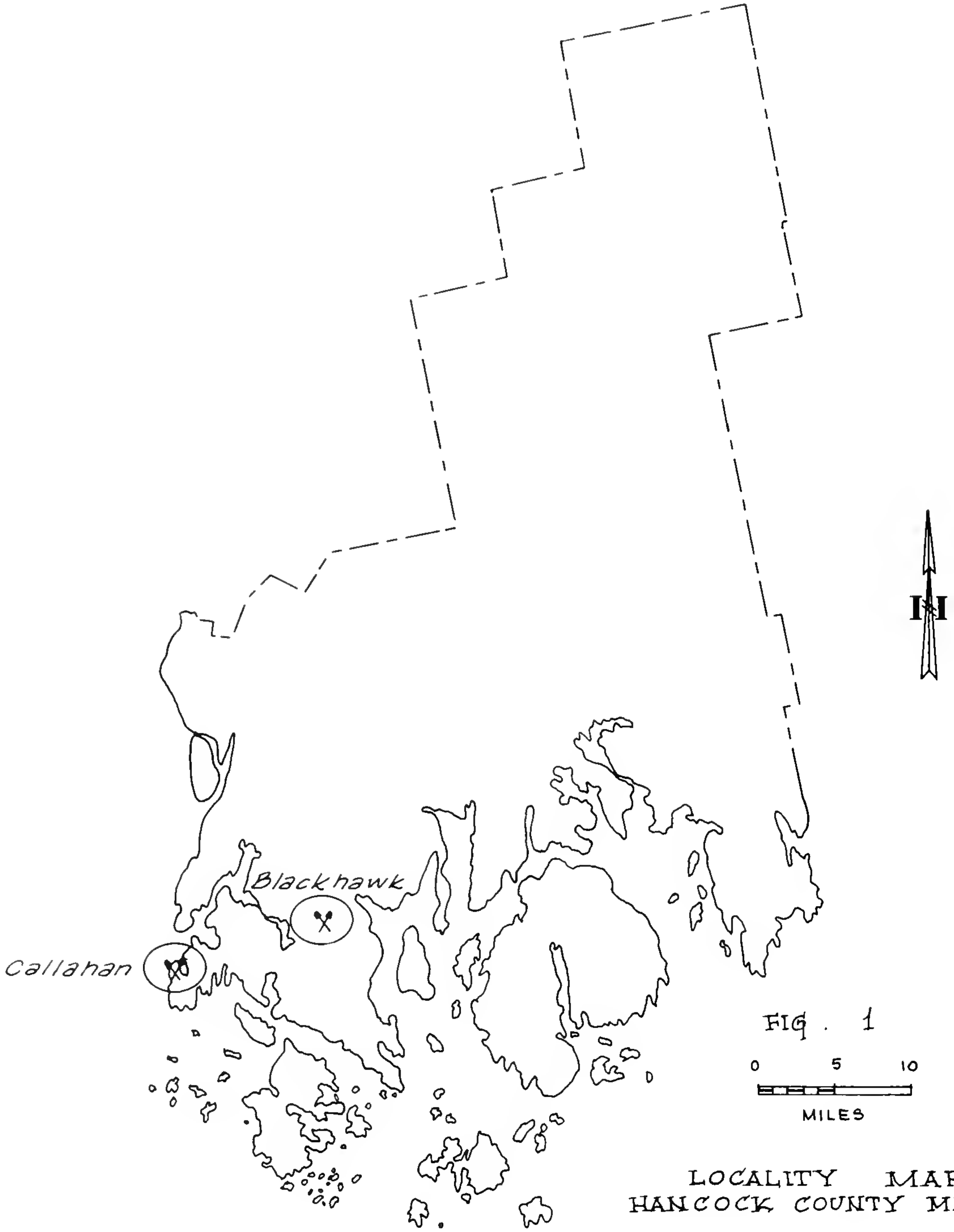
The geology and ore deposits of the area are presently under study. D.B. Stewart of the U.S. Geological Survey is continuing work in the Castine 15' minute series quadrangle and the islands to the south and west. E.S. Cheney has just completed work on geology of the Blue Hill-Castine District. Previous work includes a prospect evaluation of the Hancock County deposits by R.S. Young in 1962, and P.S. Wingard studied the area as an unpublished Ph. D. thesis (Univ. of Illinois) in 1961. Maine Geological Survey personnel have conducted numerous field studies in the area during the last decade.

The author wishes to acknowledge the assistance of staff members of the Callahan Mining Co. and Blackhawk Mines, ltd. for their assistance in obtaining permission to visit the properties and offering technical assistance.

General Geology

The area is underlain by a 30 mile wide belt of Ordovician(?) to Lower Devonian age folded and faulted metasedimentary and metavolcanic rocks, including meta-fragmental volcanic rocks. That were intruded by plutonic rocks of the Bays-of-Maine complex, mostly mafic in character, and Devonian age granite.

Although an absolute stratigraphic section has not been established, there are indications of the following series of rocks:



LOCALITY MAP
HANCOCK COUNTY MAINE

	Granitic intrusion with mafic dikes	(Acadian Orogeny)
Devonian		
Silurian	Castine formation	Metasediments and felsic metavolcanics.
	Ellsworth schist	Metasediments, quartzite metavolcanics.
Ordovician (?)		
	Penobscot black slate	Black, sulphitic slate, minor metasandstone.

Early Devonian (Acadian) regional metamorphism developed many southwesterly plunging folds. This event was followed, according to Cheney, "with the generation of northeasterly and northerly trending faults of moderate displacement that cut all of the metasedimentary units." Contact metamorphic aureoles have altered and deformed the surrounding metasediments. As a result of intrusion of the porphyritic, leucocratic quartz monzonite stocks.

With the exception of the massive replacement (?) deposit at the Harborside Mine, all of the Fe-Zn-Cu-Pb sulphide prospects which have been examined in the district occur within the contact aureoles surrounding the granitic plutons, (from Cheney, 1969).

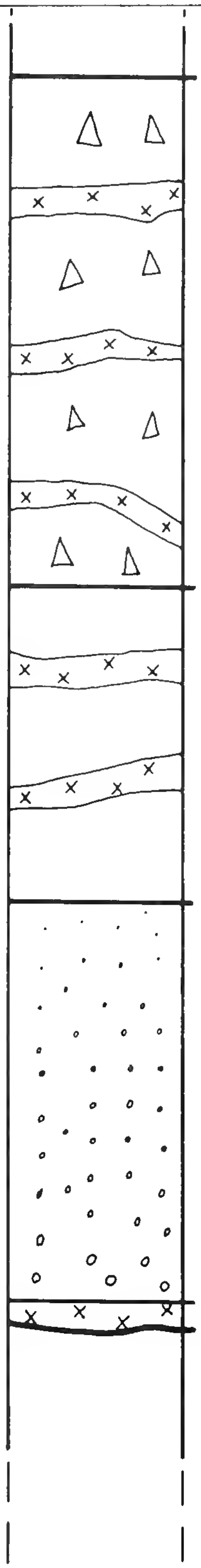
Description of Deposits

The two deposits to be visited on Trip F-3 are presumed to occur in radically different enviornments. The Harborside Mine at Cape Rosier is producing from a disseminated, massive replacement suite of sulphide minerals occurring within a section of silicious, highly altered and sheared metavolcanic rocks of the Castine formation. Cheney (1969) describes the Blackhawk Mine in Blue Hill Township, "the stratabound copper and zinc deposits....occur within a quartzite-bearing interval of the Ellsworth schist next to the northern margin of the Sedgwick pluton.

Harborside Mine. This property is located on the northshore of the Cape Rosier Peninsula close to Penobscot Bay (Figure 2). It is an open pit and adit operation, producing 750 tons per day of copper-zinc ore. A concentrating plant is located on the property, with separate zinc and copper concentrates shipped to contract smelters.

Pyrite, chalcopyrite and sphalerite are the ore minerals, occurring within a talc-chlorite-carbonate unit in the Castine formation. This unit is 120'-200' thick, wedging out (?) with depth. Figure 2 shows the stratigraphic relations of the mine area. The section strikes N.N.E. and dips S.E. at a moderate angle. The talc ore zone dips 45° at the surface, shallowing to 20°-25° with depth. The sulphides are present as fine,

F-3
4



Fragmental: Various fragment types and sizes in a siliceous matrix

Hanging Wall rhyolite breccia: angular rhyolite fragments in a chlorite to rhyolite matrix. Cut by numerous fine grained diorite dikes Average thickness of zone : 325 feet

Talc - chlorite - carbonate ore zone : sphalerite - chalcopyrite - galena. Cut by fine grained diorite dikes Zone varies from 120'-200' thick

Fragmental : rounded to angular fragments of various types and sizes in a chloritic matrix. Smaller fragments at the top. Zone varies from 150'-250' thick

Course grained diorite

Fragmental : various fragment types and sizes in a siliceous matrix.

PENOBSCOT MINE - GEOLOGIC COLUMN
FROM CALLAHAN MINING CORPORATION
GEOLOGY DEPARTMENT

dip varies from 45°-20°

SCALE: 1" = 100'

1970

fracture-filling stringers, discrete grains and massive replacement ore bands up to tens of feet wide. Most ore bands contain both zinc and copper ore, but the principal main pit ore band is predominately sphalerite.

The hanging wall of the ore zone is an extremely tough, light greenish gray rhyolite-rhyolite breccia, a common unit in the upper one-half of the Castine formation. The ore boundary is quite sharp at the breccia-talc carbonate. There is also a sharp contact with the presumed underlying dark colored, dense diorite. This latter unit composes the 'footwall' ore cutoff.

The present pit is approximately 600' in diameter, 200' depth, with a T.D. or 350' estimated. An adit at the 180' level in the northwest wall of the pit has exposed a high-grade copper-zinc stringer. This stringer is being mined as a modified cut and fill, shrinkage operation.

The geology and ore control relations are still not fully understood at this time. A very complicated series of faults and fractures exists. The geologic staff at Harborside have prepared mine models which will in part be available to the conferees. Company interests will confine the discussion relating to the property during our visit.

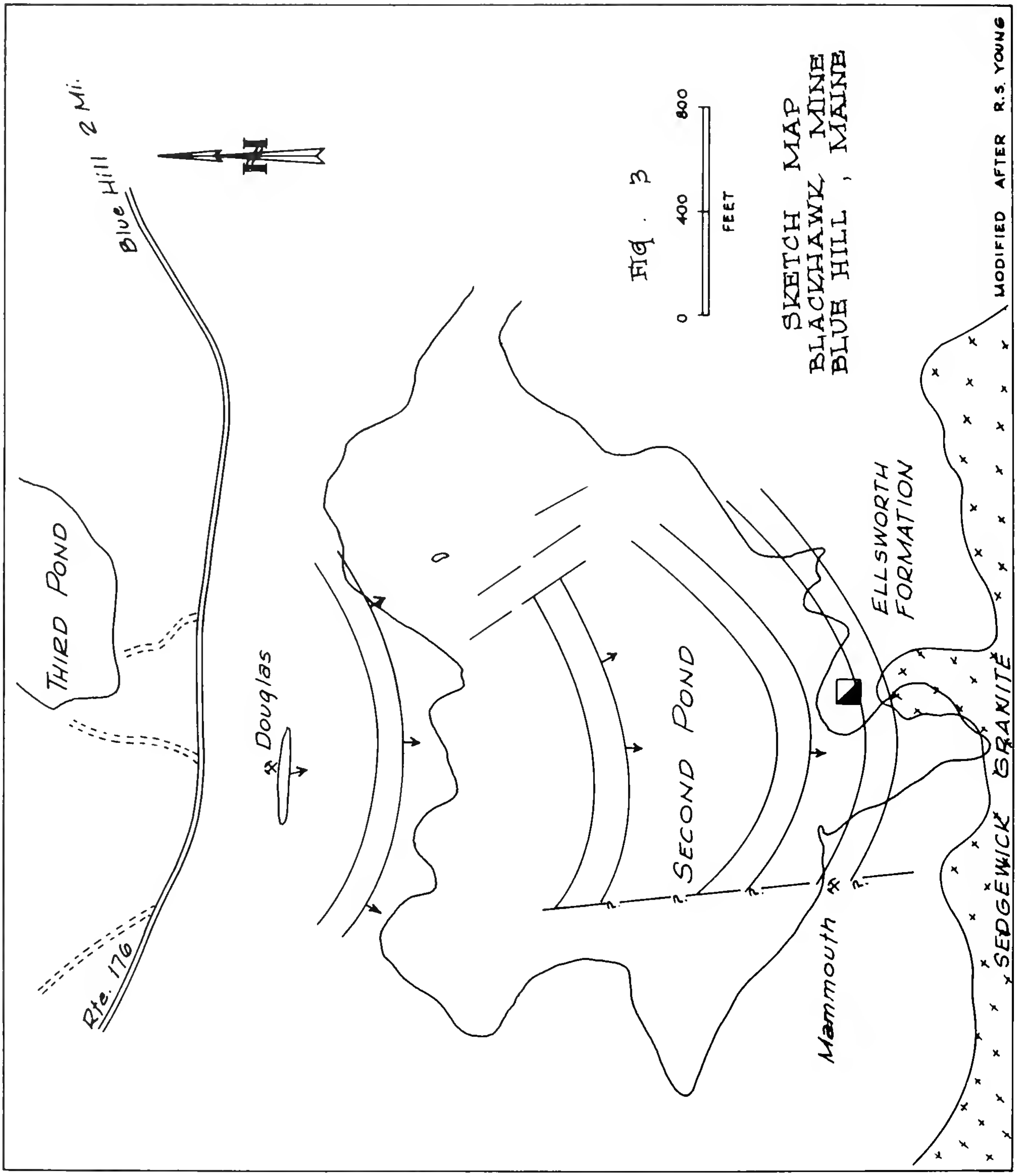
Blackhawk Mine. This property is under supervision of Denison Mines, Ltd. of Toronto. It is located 2 miles west of Blue Hill village and is composed of the grouping of several old mining operations under and adjacent to Second Pond (Figure 3). Included are the Mammouth Mine, the Douglass Mine, and the Boston Bay, Twin Lead and Pond prospects. These prospects and mines we operated a century ago, during the Maine mining boom of the 1870's to '90s. In the early 1960's, an 800' shaft was sunk to the main ore zones. Several thousand feet of laterals were driven to develop the ore. Such development ore has been stockpiled. The mine is presently flooded and on stand-by status.

The ore zones lie within a series of gently folded beds in a south plunging fold of the Ellsworth formation. These zones are confined to quartzite beds with internal red-brown biotite laminae. Ore minerals: sphalerite pyrite-chalcopyrite-occur as tabular bands and irregular but elongate pods in the sandy and biotite layers. There are three quartzite-biotite bands in the mine area. Highest zinc values occur in the biotite schist close to thick quartzite beds. The main zinc zone presents pinch and swell 'Boudin' features but maintains a fairly regular pattern around the 'Second Pond' fold.*

The Second Pond fold has an amplitude of 1700'. It is an antiform, plunging southward at 25°-30° into the irregular contact with the Sedgwick pluton. Cheney (1969) provides a genetic relationship between the pluton and the metal sulphides. The open file report at the Maine Survey office is the best description of the mine area.

---

* New term


SKETCH MAP
BLACKHAWK MINE
BLUE HILL , MAINE
Fig. 3
0
400
800
FEET
MODIFIED AFTER R.S. YOUNG
Blue Hill 2 Mi.
N
THIRD POND
Rte. 176
Douglas
SECOND POND
Mammouth
ELLSWORTH FORMATION
SEDGEWICK GRANITE

The author has noted several localities underground where a green-colored coarse phaneritic igneous rock, similar in texture to the Sedgwick pluton, contain pyrite and chalcopyrite, as irregular stringers and blebs.

TRIP SCHEDULE

Depart Rangeley	0645 hours
Arrive Bucksport	1015 hours
Depart Bucksport	1025 hours
	Outcrops of rusty weathering Penobscot formation just east of Bucksport.
Arrive Harborside	1120 hours
	Outcrops of Castine formation occur from S. Penobscot to mine.
Callahan Mine visit	1120-1400 hours
	Lunch time in ore pit.
Depart Harborside	1410 hours
Arrive Blackhawk Mine	1430 hours
Mine area examination	1430-1545 hours
Additional outcrop and old pit examination	1545-- hours
B.Y.O.D. supper at South Blue Hill	1730 hours

REFERENCES

Chapman, C.A., 1962, Bays of Maine igneous complex: Geol. Soc. Amer. Bull., V. 73, p. 883-888.

Cheney, E.S., 1969, Geology of the Blue Hill-Castine mining district, southwestern Hancock County, Maine: Maine Geol. Surv., open file report (in press, 1971), Augusta, Maine 148p-and plates.

Doyle, R.G., (as editor) with others, 1967, preliminary geologic map of Maine: Maine Geol. Surv., Augusta, Maine.

Hussey, A.M. II, and Austin, M.B., 1958, Maine metal mines and prospects: Maine Geol. Surv., Min. Res. Index No. 3, 53p.

Wingard, P.S., 1961 Geology of the Castine-Blue Hill area, Maine: unpublished Ph. D. dissertation, University of Illinois, 147p.

Young, R.S., 1962, Prospect evaluations, Hancock County, Maine: Maine Geol. Surv., Special Econ. Studies Ser. No. 2, 113p.

Glacial and Postglacial Stratigraphy Along Nash Stream, Northern New Hampshire [1]

By Fred Pessl and Carl Koteff
U.S. Geological Survey, Boston, Massachusetts

Introduction

Late in the morning of May 20, 1969, the earthfill and timber crib dam at Nash Bog Pond burst, and the erosion caused by the water rushing down Nash Stream resulted in several remarkable exposures which are the principal subject of this field trip. Nash Stream is about 8 miles long and averages less than 400 feet wide in the reaches where erosion was most extensive. Mean annual discharge, measured at the junction of Nash Stream and the Upper Ammonoosuc River, is about 90 feet^3/second. Immediately before the dam burst, heavy rain and snowmelt were sufficient to produce a flow of about 1,200 feet^3/second.[2] The major part of the flood discharge occurred in about 7 hours, although abnormally high readings were recorded for several days.

Preliminary study of the exposures along Nash Stream show the following generalized stratigraphy:

Colluvium.--One to five feet thick; ranges in texture from poorly sorted sand and granule gravel to a very poorly sorted mixture of silt- to boulder-sized fragments; locally overlies peat and wood fragments.

Upper outwash.--Maximum exposed thickness 25 feet; chiefly composed of horizontal to gently south-dipping stratified glacial-stream sediments; texture variable, ranging from moderately to poorly sorted sand and pebble gravel with minor interbeds of medium- to fine-grained sand and silt, to very poorly sorted boulder-cobble gravel. Commonly appears light gray (2.5Y 7/2)[3] to light olive gray (5Y 6/2) in exposure; fine-grained sand and silt coat the larger clasts; locally it is so compact that digging is very difficult. Locally forms terraces 10 to 50 feet above present stream level, and at one exposure appears to interfinger with the upper till.

---

[1] Publication authorized by the Director, U.S. Geological Survey. Work done in cooperation with the State of New Hampshire, Department of Resources and Economic Development.

[2] Much of the hydrological data was supplied by Charles Hale, U.S. Geological Survey, Concord, N.H.

[3] Munsell Color Company, 1954, Munsell Soil Color Chart: Baltimore, Maryland, Munsell Color Co., Inc.

Upper till.--Maximum exposed thickness 93 feet; moderately compact to friable; grain-size distribution varies greatly, ranging from more than 75 percent sand and less than 5 percent clay in inferred ablation facies, to less than 40 percent sand and more than 20 percent clay in some basal parts where mixing with fine-grained water-laid sediments from underlying units has occurred. Most common texture is 55 to 65 percent sand, 25 to 35 percent silt, and 5 to 10 percent clay; locally interbedded with water-laid sand, silt, and gravel layers. Matrix color ranges from light olive gray (5Y 6.5/2) to olive (5Y 5.5/3); strong iron-stained zones and green-gray reduced zones are locally well developed.

Lower outwash.--Maximum exposed thickness 75 feet; composed of glaciolacustrine sediments consisting of deltaic beds of very clean medium to coarse-grained sand with minor interbeds of small pebble gravel and silt and fine-grained bottomset beds of fine-grained sand, silt, and clay; large-scale bedding dips gently south; collapse structures such as faulted, tilted strata and slumped bedding are locally well developed.

Lower till.--Maximum exposed thickness 70 feet; moderately to very compact, commonly massive, locally has subhorizontal textural layering, and subhorizontal and subvertical joints; grain-size distribution ranges from 40 to 70 percent sand, 25 to 35 percent silt, and 5 to 25 percent clay; matrix color dark olive gray (5Y 2.5/2), locally iron stained to depth of 19 feet.

Tentative interpretation of the Nash Stream sections indicates that deposits of two separate ice sheets are present. The lower till and lower outwash record the presence and disappearance of the first ice sheet; the upper till and upper outwash record the advance and disappearance of the second ice sheet.

Buried wood and peat have been collected from several exposures along Nash Stream but, unfortunately, none have been collected from exposures in which the stratigraphy is firmly established. Peat overlying till that is structurally and texturally similar to the lower till and that is overlain by colluvium, has been dated at 10,200±350 years B.P. (U.S. Geological Survey sample number W-2431).[4/] Wood collected from what was initially interpreted as lower till has been dated at 8,580±300 years B.P. and 8,500±300 years B.P. (W-2430) (W-2434). In view of these later dates, we suggest that the wood-bearing deposit is probably disturbed till which was involved in a landslide.

The significance of the exposures along Nash Stream lies in the completeness of the stratigraphic section and in the similarity of the tills here with the tills described from elsewhere in New England. Direct regional correlation remains tentative, but we suggest that the

---

[4/] All age determinations by Meyer Rubin, U.S. Geological Survey.

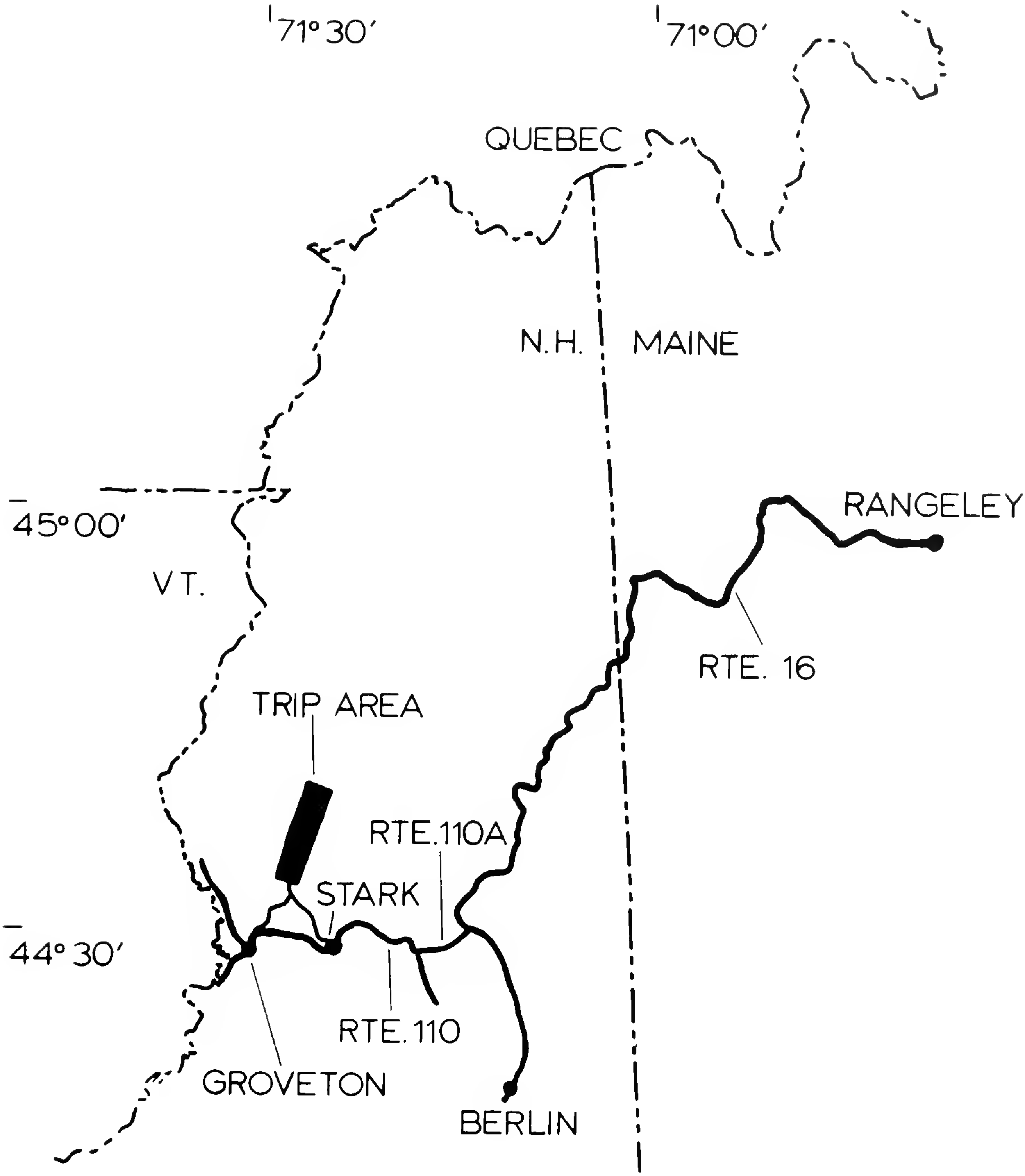


Figure 1. Index map of field-trip area, Nash Stream, N.H.

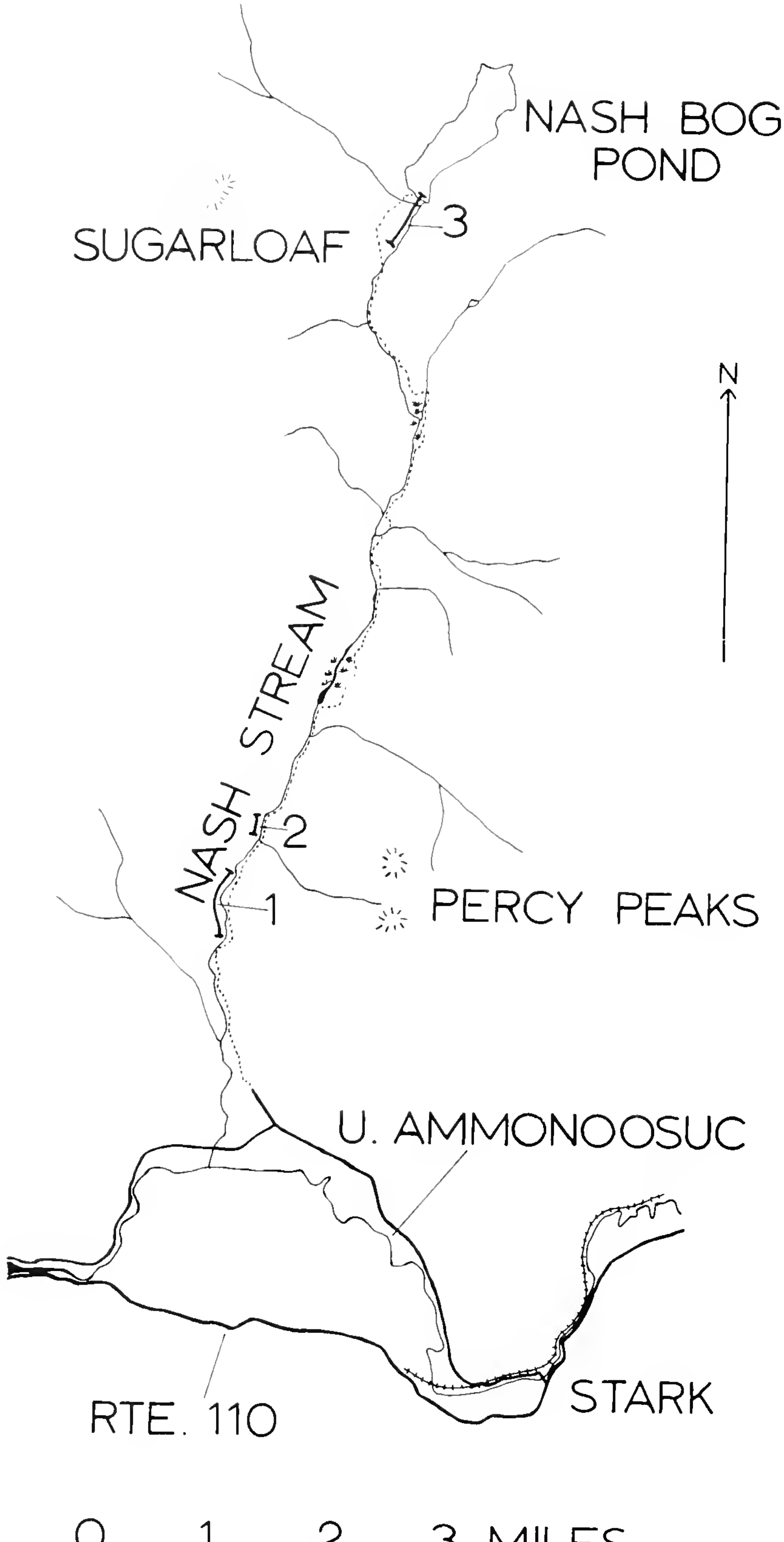


Figure 2. Map of the Nash Stream area showing field-trip stops.

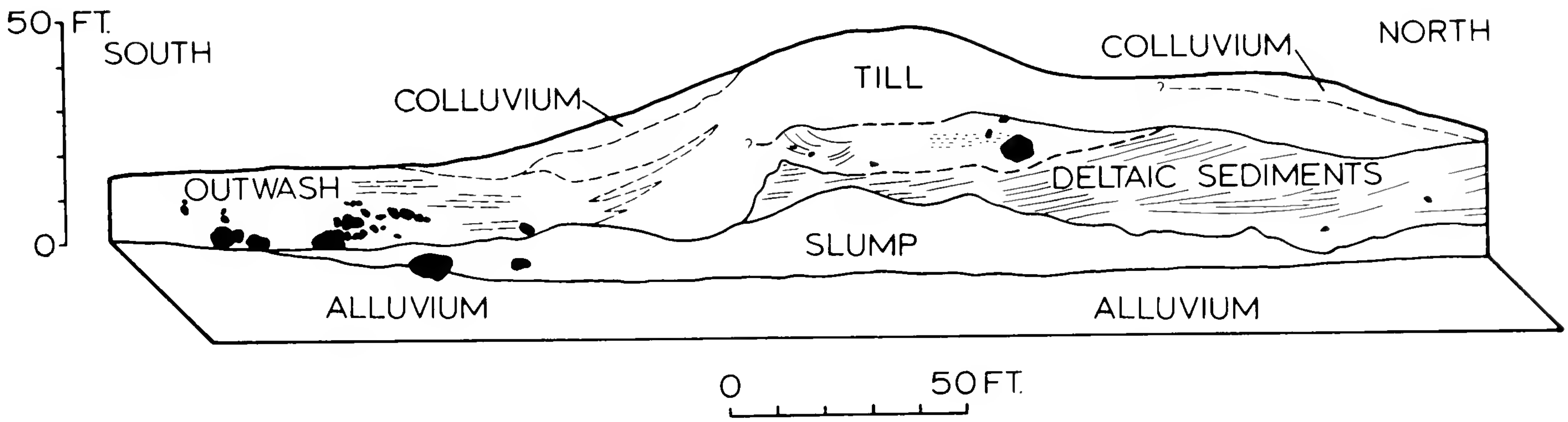


Figure 3. Generalized diagram of exposure along west bank of Nash Stream at southern part of Stop 1. Basal part of till in center of exposure is composed of deformed till and disrupted stratified sediments. Prominent boulders shown in black.

stratigraphy exposed along Nash Stream further supports the hypothesis that deposits of two separate ice sheets are widespread throughout most of New England.

The field-trip area is on the Percy topographic quadrangle map. Nash Bog Pond extends north into the Dixville quadrangle.

Mileage	Road Log
0.0	Road log begins at the intersection of State Route 110 and the access road at Stark village (fig. 1). Turn right at Union Church, cross the Upper Ammonoosuc River over Covered Bridge #37 (fig. 2). Note retired fire warden Myron Osgood's house on the right.
0.1	Turn left at intersection. Note Devils Slide on the right.
1.0	Cross Canadian National Railroad track.
4.0	Turn right at intersection of road to Sugarloaf Mountain Lookout Station. Caution: from here on the road is private; it belongs to the Groveton Paper Company, and it is travelled by logging trucks.
5.1	Gravel exposure on the right is tentatively identified as an upper sand and gravel terrace composed of outwash from an earlier glaciation, and a lower coarser gravel terrace of outwash from a later glaciation.
6.0	STOP 1: This is the major stop of the trip and includes inspection and discussion of three large exposures west of the road across Nash Stream. The first exposure is at the southern end of Stop 1 (fig. 3); it shows southward-dipping deltaic sand and gravel that lap on and overlie till which is similar to the lower or "old" till recognized in southern New England. The following section was measured in fall 1969 immediately north of the north end of the area shown in figure 3.

Thickness (ft.)	Description
5 to 5.5	Poorly sorted, very fine to very coarse grained sand and fine granule gravel with scattered pebbles and cobbles; pervasive iron staining in upper 17 inches: discontinuous strong iron and manganese staining in lower part of unit. Stratification parallels the slope and appears to be continuous with the colluvium that contains wood fragments near the middle part of the exposure below.

Thickness (ft.)	Description
2.5	Poorly sorted coarse to medium-grained sand mixed with sandy till-like material.
2.7	Very sandy olive till (5Y 4.5/3) with prominent granule-sized clasts in matrix.
1.3	Horizontally bedded, silty olive-gray clay (5Y 4.5/2) with minor interbeds of very fine grained sand. Prominent iron staining on bedding planes.
19.0	Sharp contact with fine-grained sand layer 1 inch thick overlying till. Compact blocky olive-gray till (5Y 4/2.5) with conspicuous iron-stained joint facies; two prominent joint sets, one subvertical, one subhorizontal.
	Sharp contact between this iron-stained till (above) and dark-gray nonstained till (5Y 3.5/1); at contact iron-stained surfaces of joints penetrate downward into nonstained till. Immediately above contact, a strongly iron-stained, fine to very coarse-grained sand layer 2 inches thick strikes southwest and dips northwest into slope face.
29.0	Compact, massive till: matrix color grades downward from dark gray (5Y 3.5/1) to very dark gray (5Y 2.5/2); irregularly distributed, discontinuous, subhorizontal textural layering of fine to very fine grained sand and silt is locally cut by subvertical joints.
33.0	Slump.

Gray, sandy till overlies the south-dipping deltaic beds (fig. 3) described above. This till is moderately compact to friable and locally is interbedded with fluvial sand and gravel. The basal part of the till is locally mixed with fine-grained material from the underlying water-laid sediments; deformation structures in the basal part of the till are oriented in a southerly direction. Interbedded with and overlying this till is a boulder-cobble gravel which extends southward in the form of a low terrace 10 to 15 feet above present stream level. A thin layer of colluvium overlies the terrace gravels and laps upslope onto the till. Immediately east of the road (opposite side of Nash Stream), this coarse bouldery gravel disconformably overlies deltaic or lacustrine sediments and is, in turn, overlain by colluvium which truncates the stratification. Waterworn bedrock is also exposed east of the road.

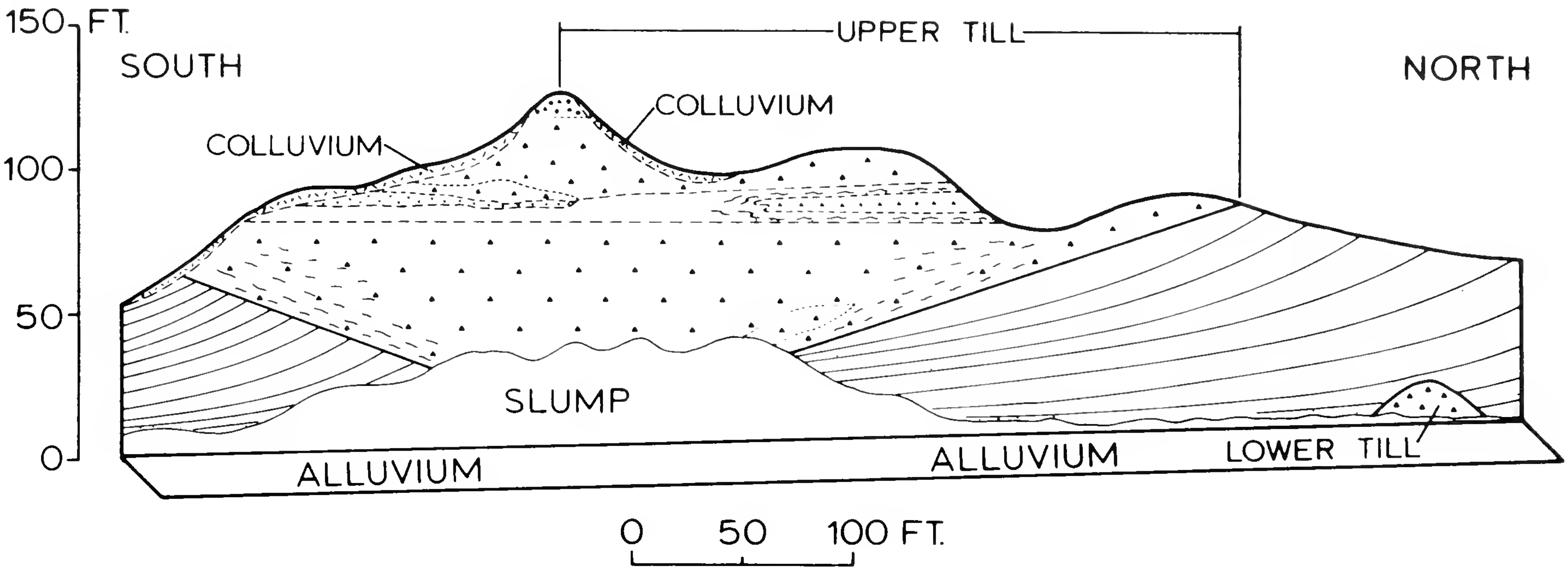


Figure 4. Generalized diagram of exposure along west bank of Nash Stream at Stop 1. Upper till includes interlayered water-laid sand, silt, and gravel; in central part of exposure interlayered sediments are strongly deformed upper 10 to 15 feet of upper till is very poorly sorted sandy bouldery till. Short dashed lines near base of upper till at northern contact and above sand and gravel interbeds near center of exposure indicate zones of inferred chemical weathering.

More than 90 feet of light olive-gray to olive-brown till with minor stratified interbeds overlies deltaic sand and gravel at the second exposure of Stop 1 (fig. 4). At the time of this writing only the following generalized description based upon reconnaissance and preliminary laboratory analyses was available.

Colluvium.--Several feet thick, overlying deltaic sands to the south and lapping onto the upper till, but apparently absent from the uppermost part of the section; lower contact subparallel to the topographic surface.

Till.--Ten to fifteen feet thick, very coarse, bouldery, and sandy; inferred to be an ablation deposit.

Till.--Twenty to thirty feet thick; sandy, moderately compact to friable, with contorted layers of well-sorted sand and silt. Contains 70 to 75 percent sand, 20 to 25 percent silt, and 5 percent clay-sized particles.

A lens-shaped zone of green-gray till surrounded by a rim of strong iron staining occurs near the base of this unit. Directly beneath this zone of apparent chemical weathering is a lens-shaped body of pebble-cobble gravel.

Stratified sediments.--Five to fifteen feet thick, composed predominantly of crossbedded, horizontally stratified, coarse to fine-grained sand with scattered pebbles and cobbles and locally intermixed with lenses and stringers of till. The horizontal stratification appears to truncate the north-dipping layering in the till beneath. Northward along the face, this stratified unit thickens to include a wedge of till and disrupted beds of fine-grained water-laid sediments.

Till.--Maximum exposed thickness 50 feet; moderately to very compact; contains prominent stringers and lenses of strongly iron-stained, coarse to very coarse grained sand; finer grained material is less prominently stained, producing a yellow-brown and olive-gray mottling in the till matrix. Grain-size distribution in the till matrix is 55 to 60 percent sand, 30 to 35 percent silt, 5 to 10 percent clay. A zone of gray (reduced?) till similar to the till lens noted in the till above occurs near the base of this unit. Faint layering (texturally controlled?) is present in the till near the contact with the underlying deltaic sands. This layering appears to be truncated by the stratified unit above but parallels the lower contact which defines the partly obscured troughlike configuration of the till.

Lacustrine sediments.--Maximum exposed thickness 75 feet; medium to coarse-grained sand with minor fine-grained sand, silt, pebbly sand, and small pebble gravel; bedding dips south 15° to 18° near upper part of unit, flattens to nearly horizontal with depth; stratification is truncated and distorted at the till contact, and shear structures produced by apparent south-directed forces are common. Admixing of the till with water-laid sediments is prominent in the contact zone.

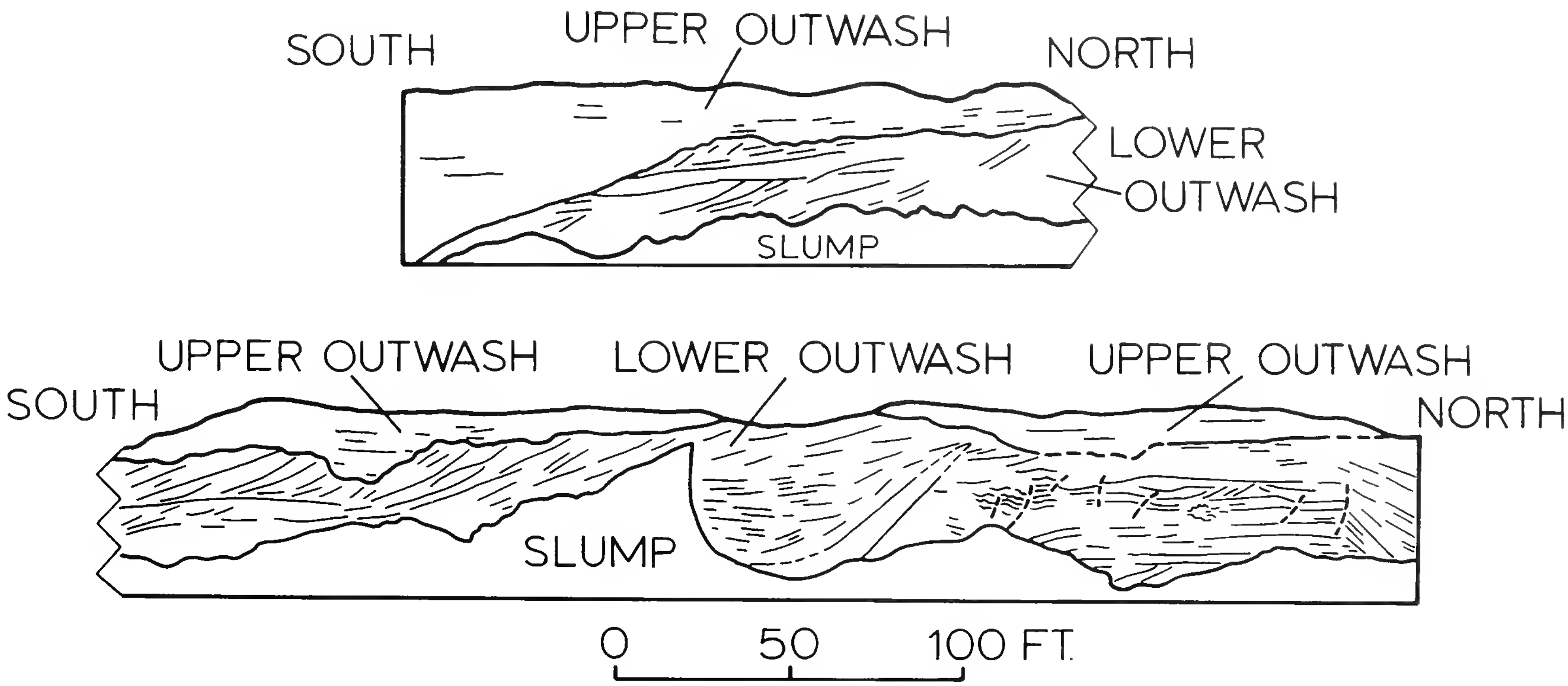


Figure 5. Generalized diagram of exposure on west bank of Nash Stream at Stop 2. Upper section extends south of the south end of the lower section. Missing interval between the two sections is about 200 feet. Vertical scale approximately equal to horizontal scale.

Farther north along this exposure, the lacustrine beds overlie another till which is exposed in two places near present stream level. The texture of the till matrix at these localities is somewhat coarser (60 to 70 percent sand, 25 to 30 percent silt, 5 percent clay-size particles) than that of the lower till exposed near the southern part of Stop 1, but the stratigraphic position of the till here clearly indicates that it predates the upper till. The upper outwash is missing at this exposure.

6.5 The third exposure of Stop 1 shows upper till overlying deltaic or lacustrine beds of sand, gravel, and silt. At the south end of this exposure, terraced gravels of the upper outwash truncate and overlie south-dipping foreset beds of the lower outwash. Further north along the exposure, till overlies apparent topset beds of the lower outwash, and still further north the till overlies and appears mixed with lacustrine silt and very fine grained sand.

6.7 To the west, upper outwash composed predominantly of cobble gravel overlying the finer grained lower outwash sand and gravel is exposed.

7.0 STOP 2: Waterworn bedrock east of road.

This exposure (fig. 5) about 1,200 feet long, shows upper outwash composed predominantly of boulder, cobble, and pebble gravel disconformably overlying conspicuously collapsed deltaic sand and gravel which is inferred to be the stratigraphic equivalent of the deltaic or lacustrine sands and gravels at Stop 1. At the southern end only upper outwash is exposed. It is very poorly sorted, tightly packed, brownish-gray bouldery gravel with many angular and subangular larger clasts. Major bedding planes dip very gently south. Three hundred to four hundred feet north of the southern end of the exposure, the contact between the two outwash units rises about 18° to the north. Bedding in the lower outwash dips 15° to 25° S. and is truncated at the contact with the upper outwash. In one place near the middle of the exposure, a large channel, about 50 feet wide and 15 to 20 feet deep, has been cut in the lower outwash and filled by upper outwash. The lower outwash is composed of moderately to well-sorted, clean, very coarse to medium-grained sand and interbedded pebbly sand and small pebble gravel. The coarser grained sand and pebbly sand layers are light purple to pink.

Although collapsed bedding in the lower outwash occurs along most of the exposure, the most prominent collapse structures are in the northern half of the exposure. Near the northern end, alternately darker and lighter colored flat-lying layers of interbedded medium- to fine-grained sand and coarse-grained sand to small pebble gravel have been faulted and tilted, dipping about 25° N. Noncollapsed upper outwash and colluvium overlie the collapsed lower outwash. The upper outwash clearly postdates the collapse of the lower outwash.

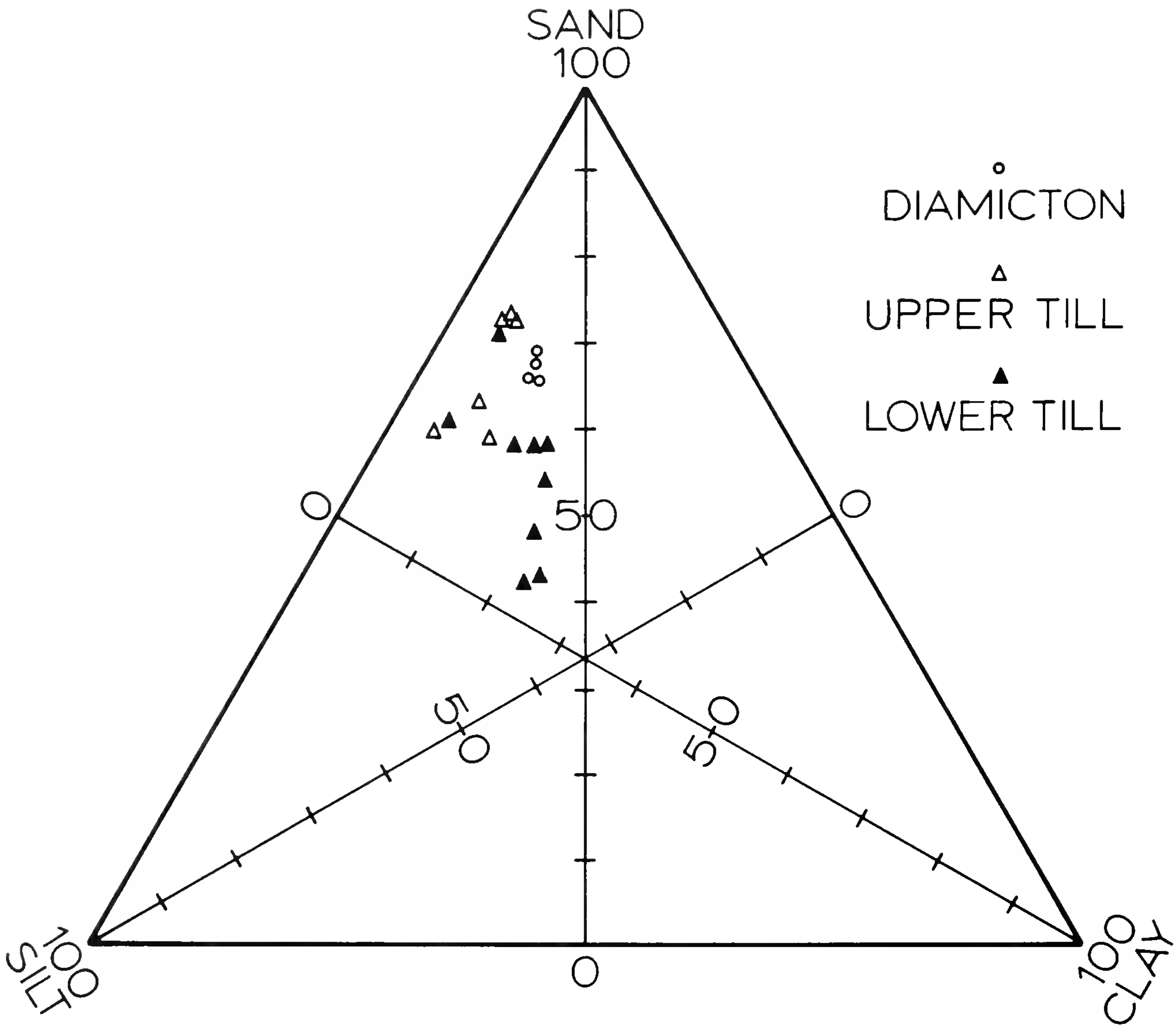


Figure 6. Grain-size distribution (in percent) in the matrix of till and an inferred landslide deposit (diamicton) from the Nash Stream area.

8.7 In a small pit to the east, about 10 feet of upper till is overlain by 1 to 30 feet of strongly iron-stained pebble-cobble gravel. The till has prominent sandy partings which have been accentuated by frost action, producing a blocky structure in the till at the pit face.

11.2 To the west is the southern end of a broad swampy area where the valley widens. Very little flood damage occurred here.

13.1 "Rowden's Roost" on left. Signs mark where Mr. Rowden spent the night after climbing the tree to avoid the flood waters when the dam burst. Rowden's companions scampered up the slopes to the east and hiked 5 miles through 2 feet of snow to the next valley.

13.7 Bridge over Nash Stream.

14.0 STOP 3: Road to damsite.

Descend north side of the dam abutment and walk east to the stream. View north across the basin of former Nash Bog Pond. At normal gate level the pond occupied about 240 acres and had an estimated average depth of 10 feet.

In the vicinity of the breached earth-fill structure, note exhumed logs in eroded pond-bottom sediments.

On the east bank, immediately below the dam, is a nearly horizontal iron-stained horizon. This horizon is continuously exposed around the rim of the eroded bottom sediments at the top of the scarp and can be traced beneath the artificial fill of the dam on the west side of the stream, indicating that the material beneath the oxidized zone predates the dam. A similar iron-stained zone occurs beneath the vegetated mat on the former pond bottom.

A log collected from the scarp face has a radiocarbon age of 8580±300 years B.P. (W-2430). The texture of the matrix material is 68 percent sand, 20 percent silt, and 12 percent clay-size particles, similar to tills exposed downstream, although it has somewhat more sand than most lower till, and somewhat more clay than most upper till (fig. 6). The texture and distribution of the coarser fragments, especially the boulders, is different from those usually found in tills in New England, but the nature of this difference is hard to define. The relatively young age of this deposit primarily influences our interpretation of it as a landslide deposit.

An alternative interpretation has been suggested: that this deposit may be undisturbed till representing a readvance which flowed out from a center in the White Mountains about 8,600 years ago. The presence of peat, dated about 8,300 years B.P. (W-2473), exposed at the surface nearby, and peat, dated 10,200±350 years B.P. (W-2431),

beneath colluvium overlying till further downstream, seems to argue against this hypothesis. Furthermore, we are unaware of any regional geological or palynological evidence that supports the possibility of a glacial readvance about 8,600 years ago.

Proceed downstream along west side of Nash Stream.

Distance south of dam (ft.)	
1,000	Dissected bog deposit exposed in west bank. Organic-rich layer, about 4 feet thick contains logs dated 8,460±300 years B.P. (W-2470) and peat dated about 8,300 years B.P. (W-2473). Directly beneath the peat is a layer of light olive-gray "till-like" material (landslide deposit?) which overlies poorly sorted strongly oxidized cobble-pebble-boulder gravel.
1,500	Cut bank exposes landslide debris(?) overlying bedrock or possibly a large boulder. The material locally shows strong iron and manganese staining. The matrix material is 69 percent sand, 20 percent silt, 11 percent clay-sized particles, and the texture is strikingly similar to the texture upstream at the damsite. The upper few feet of the deposit here appear to be frost disturbed. Directly opposite on the east side of the stream, well-preserved grooves and striations bearing S.30° to 35°E. occur on bedrock just above stream level. Olive to olive-gray (5Y 4.5/2.5) till resting directly on this striated surface contains 58 percent sand, 26 percent silt, and 16 percent clay-sized particles.
2,000	Exhumed peat exposed on surface of gravel bar in streambed. Peat occurs in basal part of 4-inch thick organic-rich fine-grained sand and silt which overlies oxidized pebble gravel. This in turn overlies dark-olive-green to olive-gray till. This peat is dated 10,380±350 years B.P. (W-2433).
2,100	Till exposure on east side of stream shows two textural facies of the dark olive-green till, a sandy facies overlying a clayey facies. The matrix of the sandy facies contains 58 percent sand, 30 percent silt, 12 percent clay-sized particles; the matrix of the clayey facies contains 43 percent sand, 34 percent silt, and 24 percent clay-sized particles. Both of these matrix-textures fall within the range of textures found in the lower

till downstream, and the till here is tentatively correlated with that lower till. Overlying the dark-olive-green till is a layer of peat 1 to 2 1/2 feet thick which is dated 10,200±350 years B.P. (W-2431). Strongly oxidized coarse gravel is locally interbedded between the peat and till. Colluvium 1 to 2 feet thick overlies the peat.

2500 Cut bank on west side of stream exposes mottled (irregularly oxidized and reduced) "till-like" material containing wood fragments. The matrix material is similar to that at the dam, containing 65 percent sand, 22 percent silt, and 13 percent clay-sized particles. Wood fragments and bark chips lining the cast of a log, 10 to 12 inches in diameter, were collected from this deposit. Farther downstream, a well-preserved log embedded about 4 feet below the surface was discovered by E.E. Lambert. This log has been dated at 8500±300 years B.P. (W-2434).

2,600 Dark olive-gray to olive-green "till-like" material is exposed at the base (near stream level) of the cut bank. This material is overlain by 6 to 10 inches of an iron-stained, mottled unit similar to that in the previous exposure. Strongly oxidized, coarse-grained sand and pebble gravel 12 to 30 inches thick with minor interbeds of fine-grained sand and silt overlies the iron-stained unit. The matrix of the nonoxidized basal unit contains 69 percent sand, 20 percent silt, and 11 percent clay-sized particles, very similar to the texture of the material at the dam. Numerous, well-preserved logs extend out from the face of the nonoxidized material.

In summary, the stratigraphy exposed north of the bridge along Nash Stream includes till on the east side of the stream equivalent to the lower till described at Stop 1 and a large landslide deposit on the west side of the stream composed of reworked till containing logs and wood fragments. Definition of these two units is based primarily on the age and position of the peat and logs and on the similarity of the till on the east side of the stream to that of lower till downstream.

The lateral extent of the landslide deposit is unknown, but the topography suggests that the road between the bridge and the damsite traverses the main slide mass.

Trip H

Nature of the Taconic orogeny in the Cupsuptic quadrangle, west-central Maine[1]

By David S. Harwood
U. S. Geological Survey, Washington, D. C.

Introduction

In recent years considerable evidence has been accumulated in the central and northern Appalachians to support the major hiatus between the pre-Silurian and Silurian rocks, proposed by Rogers (1838, p. 37) more than a century ago. Although the nature of the structural break and the length of the hiatus vary considerably over the Appalachian region, the data summarized by Pavlides and others (1968) indicate that diastrophism took place in much, but not all, of the northern Appalachians from late Middle Ordovician to Early Silurian. Hall (1969) and Zen (1967, 1968) have recently presented evidence for early Middle Ordovician tectonism which Zen (1968, p. 131) suggested was an early phase of the Taconic orogeny.

In the southeastern part of the Cupsuptic quadrangle there is an angular discordance between the upper Middle Ordovician (graptolite zone 12) rocks and the overlying Lower Silurian rocks. It is the purpose of this trip to examine the rocks and structural styles above and below the pre-Silurian--Silurian boundary in the hope of demonstrating some of the effects of the Taconic orogeny.

The trip involves two fairly long but not very strenuous hikes over old logging roads and through relatively open woods. Hiking boots are strongly recommended.

Stratigraphy

Stops will be made to look at the Albee, Dixville, and Rangeley Formations in the southeastern part of the Cupsuptic quadrangle. Brief lithologic descriptions of these formations with references to more detailed descriptions as well as discussions of their age and stratigraphic order are given in trip A-3 (Harwood, Green, and Guidotti, this volume) and trip A-1 (Moench and Boudette, this volume) and will not be repeated here. Lithologic descriptions of specific rocks seen on this trip are given under the appropriate stops in the accompanying road log.

---

[1]Publication authorized by the Director, U. S. Geological Survey

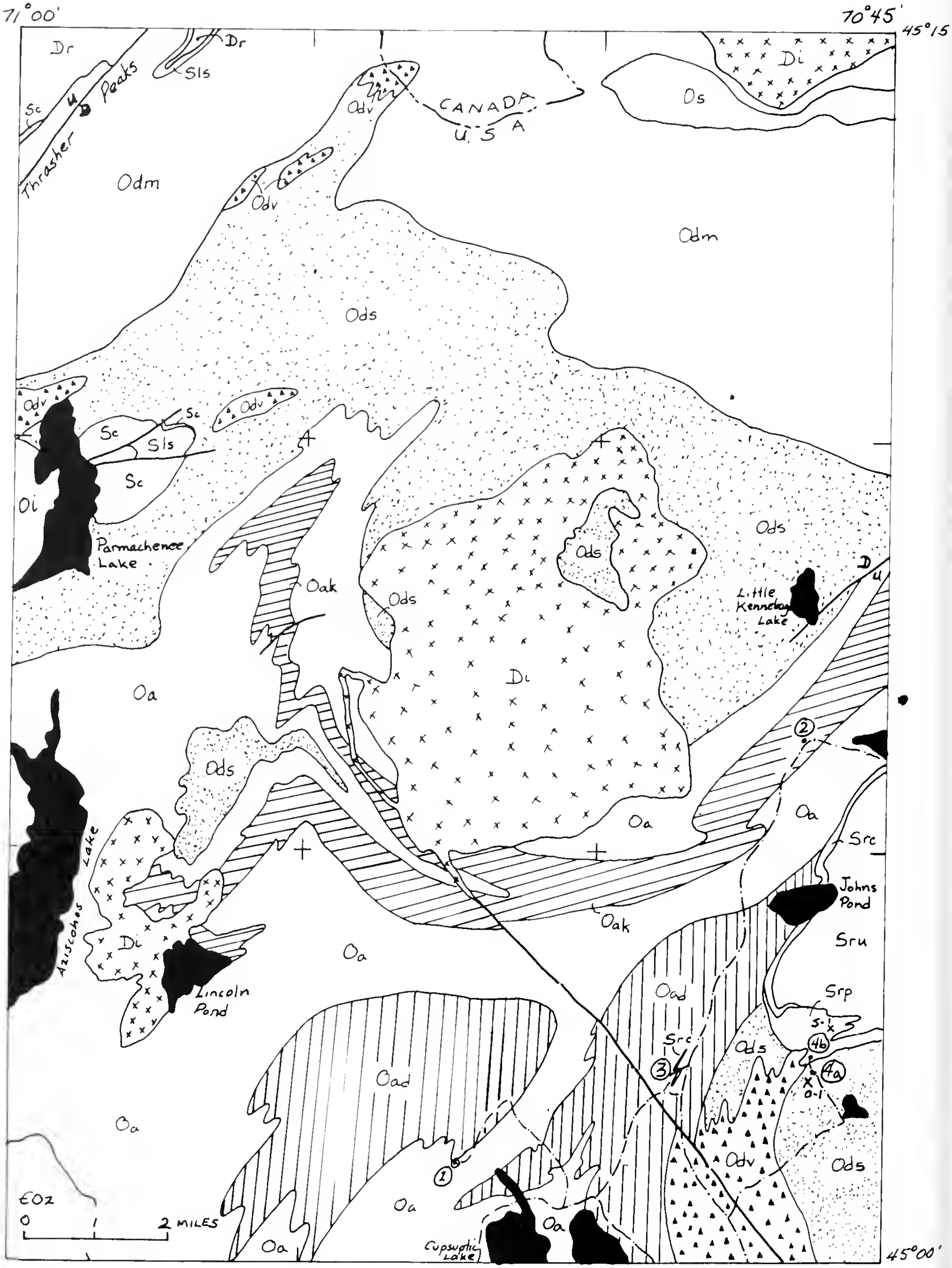


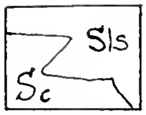
Figure 1 - Generalized geologic map of the Cupsuptic quadrangle, Maine

EXPLANATION

Dr

Gray slate and feldspathic quartzite

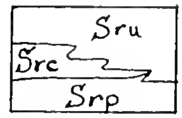
Devonian



Unnamed rocks

Sls- gray clacareous slate, limestone, and limestone conglomerate

Sc- quartz-pebble conglomerate and minor polymict conglomerate



Rangeley Formation

Sru- gray slate and feldspathic quartzite with scattered layers of polymict conglomerate

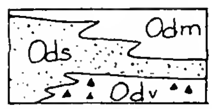
Src- quartz-pebble conglomerate and quartzite

Srp- polymict conglomerate and feldspathic quartzite

Rangeley Formation considered to Lower Silurian (?)

Silurian

UNCONFORMITY

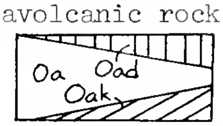


Dixville Formation

Odm- Magalloway Member; feldspathic graywacke granule conglomerate, minor slate and metavolcanic rocks

Ods-black sulfidic slate and feldspathic quartzite

Odv- mafic metavolcanic rocks



Albee Formation

Oad- green, purplish-gray and black phyllite with less than 10 percent of interbedded "pinstripe" quartzite

Oa- green to greenish-gray slate and phyllite with 10 to 50 percent "pinstripe" quartzite

Oak- red slate with greater than 10 percent of white weathering "pinstripe" arenaceous rocks

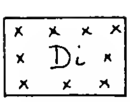
Ordovician

O€z

Aziscohos Formation of Green (1959)

Green phyllite and chlorite schist with abundant stringers and pods of white quartz

Camb. & Ord.



Devonian intrusive rocks

Oi

Ordovician intrusive rocks

Os

Ordovician serpentinite and related mafic rocks

Field trip route showing location of stops

O-1
X

Fossil locality

Fault

D,downthrown side; U, upthrown side

Pre-Silurian--Silurian boundary

Near Johns Pond (fig. 1) polymict and quartz-pebble conglomerate of the Rangeley Formation truncate the steeply dipping contact between the Middle Ordovician Dixville Formation and the underlying Albee Formation. The contact is not exposed here, but the bedding attitudes shown on figure 2 clearly indicate an angular discordance between the Silurian and pre-Silurian rocks. About 10 miles to the southeast, however, the Rangeley Formation at the type locality appears to rest conformably above the Greenvale Cove Formation which, in turn, rests conformably above the Quimby and Dixville Formation (see trip A-1, Moench and Boudette, this volume).

The pre-Silurian--Silurian boundary near John Pond has been interpreted by Harwood (1966; Harwood and Berry, 1967) as an angular unconformity because the polymict conglomerate contains abundant clasts of the subjacent pre-Silurian rocks, and there is no evidence of brittle, low-angle thrust movement in the rocks near the contact. The absence of the Greenvale Cove and the Quimby Formations is due, in this interpretation, to either nondeposition or erosion on the unconformity.

Recently Moench (1970, p. 1492) has suggested that the inferred unconformity might be a major premetamorphic fault along which wet sediments of the Rangeley Formation slid southeastward, becoming folded and cleaved in the process. This interpretation explains the absence of the Greenvale Cove and Quimby Formations and the absence of brittle fault deformation in the Silurian rocks, but it is incompatible with the fact that the pre-Silurian--Silurian boundary is folded into a northeast-trending syncline at Johns Pond. The premetamorphic faults mapped by Moench to the southeast are not folded about northeast axes; in fact, Moench's concept of down-to-basin faulting and folding requires that the northeast-trending folds be generated by slumping on unfolded premetamorphic faults.

The distribution of sedimentary facies in the Silurian rocks supports the concept of a hinge line between Johns Pond and Rangeley Lake proposed by Moench (1970) to separate a basin of apparently continuous sedimentation to the southeast from an eroding positive area to the northwest. At Johns Pond the quartz pebble conglomerate overlaps the northwestern edge of a wedge of polymict conglomerate that thickens from about 1,000 feet south of Johns Pond to about 9,000 feet near Rangeley Lake (Moench, 1970). The quartz-pebble conglomerate apparently marks a local southeastern limit of the beach environment in Llandovery time. The northwestern extent of the Llandovery beach, or, conversely the southeastern limit of the positive area in Llandovery time is not precisely known, but it must have been between Johns Pond and Thrasher Peaks (fig. 1) because the basal Silurian rocks near Thrasher Peaks are most probably Ludlow (Green, 1968, p. 1516).

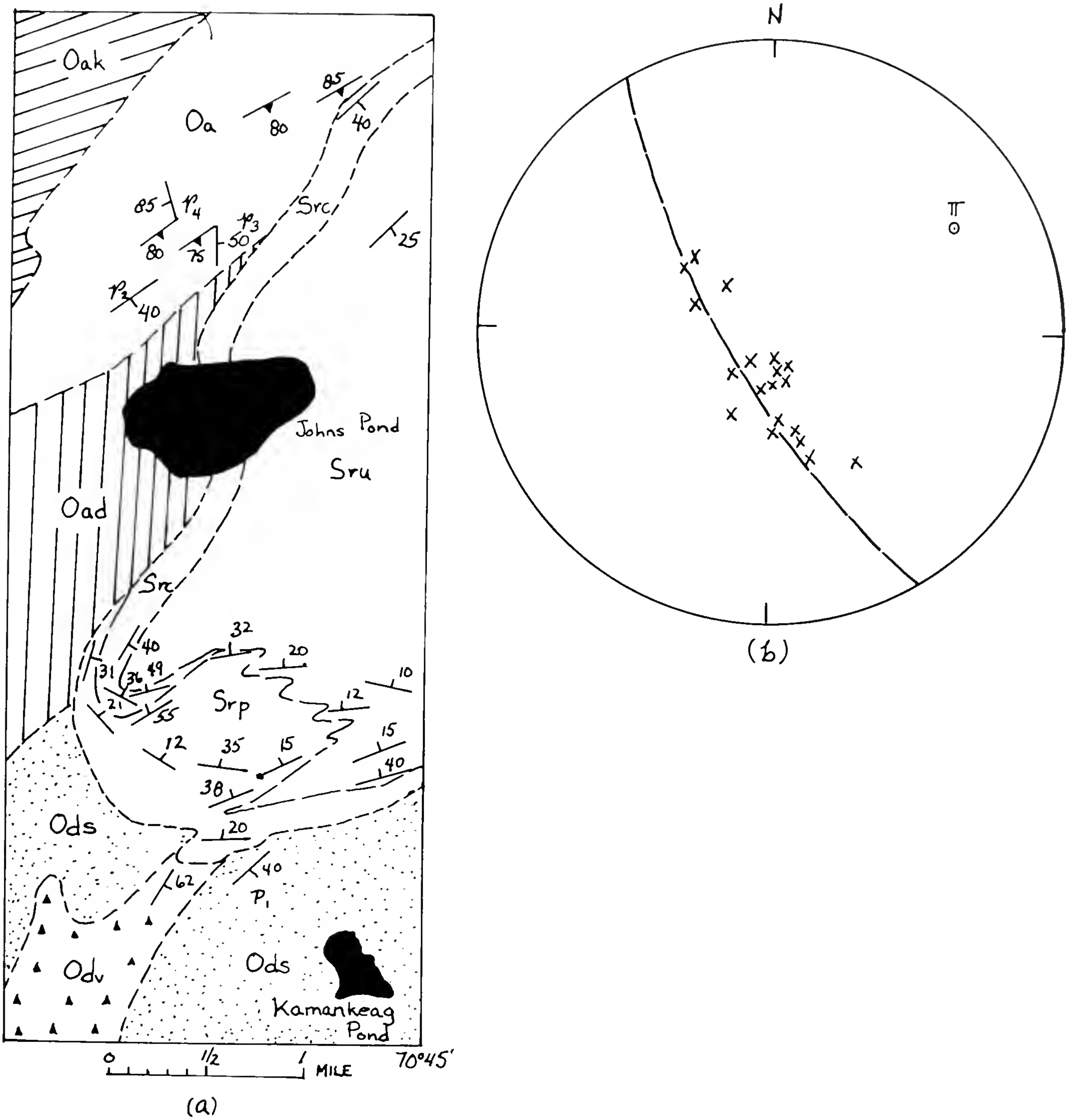


Figure 2. a) Geologic map of Silurian and pre-Silurian rocks near Johns Pond showing bedding (40) and cleavage (85).

b) Point diagram showing girdle (dash great circle) and girdle axis (π) defined by poles to bedding (x) for Silurian rocks near Johns Pond, southeastern part of the Cupsuptic quadrangle.

From these relationships it is concluded that the pre-Silurian--Silurian boundary in this area is a time-transgressive unconformity that developed on an eroding positive area underlain in part at least by pre-Silurian rocks occupying the core of the present Boundary Mountain anticlinorium. Cady (1968, p. 156) shows this positive area as the Somerset geanticline, and Boucot (1968, p. 88) shows it as Somerset Island. The onlap of the Silurian sea represents only the final curtain call to a period of diastrophism, and we logically ask, when did the show open and what were the main events?

Affects of the Acadian orogeny

All the stratified rocks in the Cupsuptic area were deformed during the Acadian orogeny in early Middle Devonian time. This deformation must be identified and effectively removed from the total deformation of the pre-Silurian rocks in order to examine the effects of the Taconic orogeny.

Bedding attitudes and graded beds indicate that the Silurian rocks near Johns Pond have been folded into a northeast-trending syncline. Poles to bedding, shown in figure 2, define a girdle whose axis plunges 18° N. 58° E. and represents the local axis of Acadian folding. The trend of this axis agrees fairly well with the trend of fold axes in the Deer Brook syncline (Harwood, 1969) to the northwest and with the trace of most of the folds outlined by Silurian and Devonian rocks in the Merrimack synclinorium to the southeast (see Osberg, and others, 1968, fig. 18-2, p. 242). Northwest-trending folds are found locally in the Silurian and Devonian rocks, but these folds appear to be local anomalies in the regional northeast trend related to deformation produced by large Devonian plutons (Guidotti, 1965).

The trend of the axial surfaces in many folds in the pre-Silurian rocks in the Cupsuptic quadrangle is between N. 40° E. and N. 50° E. and thus closely parallels the trend of the folds in the Silurian rocks. Generally, however, this is the only geometric feature that the folds above and below the unconformity have in common. Typically the folds in the Silurian rock are relatively open with gently dipping limbs, gently plunging linear axes, and simple forms. Pre-Silurian folds with a northeast trend, on the other hand, are typically isoclinal with steeply plunging commonly curvilinear axes and have complex forms.

At Stop 1 the complex fold pattern in the Albee Formation can be related to the superposition of northeast-trending folds on west-northwest-trending folds which has produced a series of small domes and basins separated by saddle-shaped areas. At Stop 2 isoclinal northeast-trending folds in the Albee have strongly curved axes resulting in "hog-back" shaped minor folds, but there is no evidence of a preexisting fold pattern or of a folded cleavage. In fact, most of the axes of the minor folds in the pre-Silurian rocks are steeply plunging or vertical, and these folds show no refolded linear or planar features.

The different fold patterns above and below the unconformity suggest that the Acadian folds were superimposed on previously folded beds, but the absence of an early pervasive flow cleavage argues against intense deformation in pre-Silurian time.

Pre-Acadian attitude of the pre-Silurian rocks

The pre-Acadian attitude of the pre-Silurian rocks near Johns Pond can be determined graphically by unfolding the Silurian rocks about the axis of Acadian folding shown in figure 2. Only an approximate orientation can be obtained because there is no way of accurately determining the slope of the erosion surface that developed on the pre-Silurian rocks. For simplicity, it will be assumed that the erosion surface was flat and thus comparable to the one beneath the Atlantic Coastal Plain sediments in New Jersey which dips slightly less than 1° (J. P. Minard, 1970, pers. comm.).

The attitude of bedding in the Dixville Formation near fossil locality 0-1 (fig. 1) rotates from N. 43°E./40°SE. (p_1, fig. 3a) to N.62°E./62°SE. (p_1^1, fig. 3a) when the Acadian axis of folding is restored to the horizontal and the overlying Silurian rocks are rotated to the horizontal. Greater variation is shown in the bedding attitudes in the Albee near the Silurian rocks north of Johns Pond; the present and pre-Acadian attitudes of bedding for three localities are shown in figures 3b, c, and d.

Although this specific geometric restoration cannot be applied throughout the Cupsuptic area because the sense of rotation must be obtained from immediately adjacent Silurian rocks, the examples in figure 3 clearly show that some of the pre-Silurian rocks were locally steeply inclined prior to Silurian time. The assumption that the erosion surface on the pre-Silurian rocks dipped a significant amount to the southeast toward the maximum thickness of the Silurian polymict conglomerate only steepens the pre-Acadian dips in the pre-Silurian rocks near Johns Pond.

Beginning of the Taconic orogeny

Zen (1967, 1968) recently summarized the evidence for Ordovician diastrophism in the Taconic area, from which he concluded that tectonic events began by block faulting, minor volcanism, and folding in earliest Middle Ordovician time. These events were followed by the deposition of Middle Ordovician carbonate rocks and black mud (graptolite zone 11 and 12 time) and eventually by emplacement of at least the early slices of the Taconic allochthon (graptolite zone 13 time). Both the allochthonous and autochthonous rocks are unconformably overlain by Silurian rocks in eastern New York. Hall (1969) also has recognized an unconformity at the base of Middle Ordovician clastic

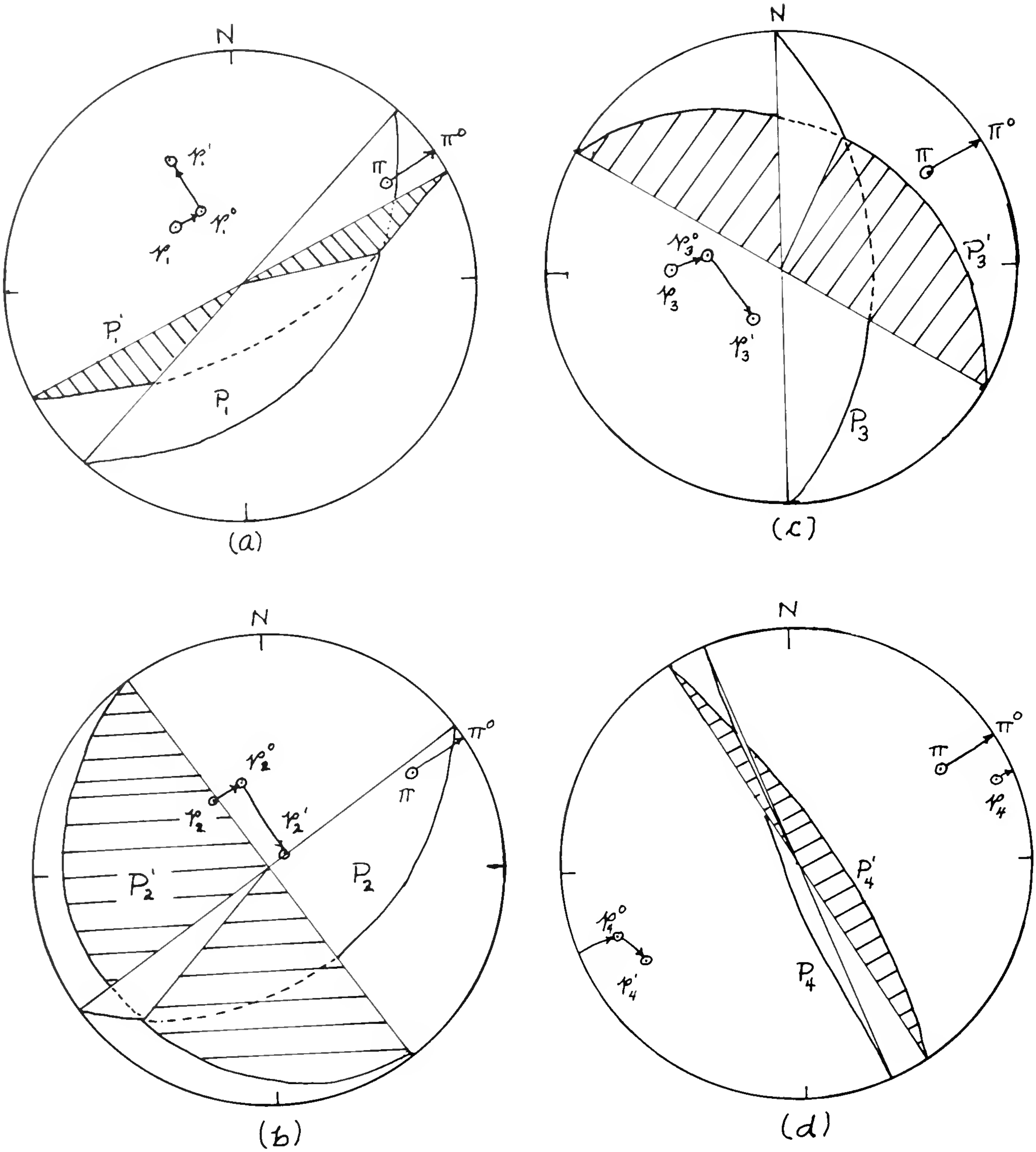

N
(a)
(c)
(b)
(d)
π
π°
p_1'
p_1°
p_1
P_1'
P_1
p_3°
p_3
p_3'
P_3'
P_3
p_2°
p_2
p_2'
P_2'
P_2
p_4°
p_4'
p_4
P_4'
P_4

Figure 3(a). Pre-Silurian orientation of bedding plane in Dixville Formation at fossil locality 0-1.

P_1 = bedding plane in present, post-Acadian, orientation

p_1 = pole to plane P_1

π = axis of Acadian folding in Silurian rocks near Johns Pond determined in figure 2.

$\pi°$ = axis of Acadian folding restored to horizontal

P_1^1 = bedding plane P_1 in pre-Silurian orientation

p_1^1 = pole to bedding plane P_1^1.

Pole p_1 rotates to p_1° as axis of Acadian folding (π) is restored to horizontal ($\pi°$) along east-west axis of the net. With $\pi°$ at the north pole of the net, p_1° rotates along a small circle of the net to p_1^1 as Silurian rocks on south limb of syncline are rotated clockwise 24° to horizontal position.

3(b), (c), (d). Pre-Silurian orientations of bedding planes in the Albee Formation adjacent to Silurian rocks north of Johns Pond. Symbology the same as in figure 3(a) except that p_2°, p_3°, p_4° rotate to p_2^1, p_3^1, and p_4^1, respectively, as overlying Silurian rocks are rotated 40° counterclockwise to horizontal.

rocks and has recognized different fold styles above and below this unconformity in north-central Maine. He concluded that folding, uplift, and erosion occurred prior to graptolite zone 11 time and again between graptolite zone 13 time and Early Silurian time.

Folding of the pre-Silurian rocks in the Cupsuptic area must have occurred after graptolite zone 12 time and before C-3-C-5 time of the late Llandovery. This corresponds very well with the second period of folding recognized by Hall in north-central Maine and with the time of emplacement of the early slices of the Taconic allochthon determined by Zen. To date, however, there is no unequivocal evidence in the Cupsuptic area for an unconformity or different fold styles within the pre-Silurian rocks which might indicate early Middle Ordovician tectonism comparable to that found elsewhere. To further complicate matters, no fossils have been found in rocks below the zone 12 graptolite-bearing slate, so there is no way of knowing whether a significant hiatus exists within these older rocks. Therefore, the stratigraphic position of the pre-zone 11 break reported elsewhere and the geologic history in this area during that time are a matter of speculation.

Hall (1969) proposed that the contact between the Aziscohos Formation of Green (1959) and Albee Formations might correspond to the unconformity he mapped beneath graptolite zone 11-bearing rocks. This possibility cannot be refuted, but certainly the most profound change in the sedimentary record below the zone 12 graptolite-bearing slate is found at the Albee-Dixville boundary. Here hematite- and magnetite-bearing pelitic rocks in the Albee are succeeded abruptly and conformably with minor interbedding by sulfide-bearing carbonaceous pelitic rocks and thick accumulations of volcanic rocks and feldspathic graywacke. In fact, the Magalloway Member of the Dixville is lithologically similar to the graywacke, granule conglomerate, slate and volcanic rocks assigned by Hall (1969, p. 468) to his Group B which contains graptolites indicative of zone 11 and 12 and which rests unconformably above pre-zone 11 rocks.

The distribution of rock types in the Dixville is compatible with uplift and erosion north and northeast of the Cupsuptic quadrangle. The thick sequence of feldspathic graywacke and granule conglomerate mapped as the Magalloway Member of the Dixville (fig. 1) may have been derived from a positive area to the north and east. The Magalloway Member is succeeded to the south by black slate containing scattered patches of mafic volcanic rocks. Along the southeastern flank of the Boundary Mountain anticlinorium a thick accumulation of mafic volcanic rocks is in contact with the Albee and is succeeded upward by black slate which, in turn, is overlain by graywacke and felsic metavolcanic rocks of the Quimby Formation.

This distribution of rocks suggests rapid deposition of the Magalloway Member from the north into a euxinic basin that was concurrently receiving large amounts of mafic volcanic rocks further to the south. It implies, furthermore, that the map units in the Dixville Formation are lithofacies, an implication supported by some fossil evidence. Neuman (1968, p. 40) equates brachiopods found by E. V. Post in the Brassua Lake quadrangle in feldspathic graywacke (see U. S. Geol. Survey, 1961, p. A12) similar to the Magalloway Member with the zone 12 graptolites found in black slate at locality O-1 (fig. 1).

It is tentatively concluded, therefore, that the early Middle Ordovician tectonism marked by an unconformity at the base of zone 11 graptolite-bearing rocks in the Taconic area and in north-central Maine is marked by the conformable change from Albee-type sedimentation to Dixville-type sedimentation in the Cupsuptic area.

References

Billings, M. P., 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. America Bull., v. 48, no. 4, p. 463-566.

________________ 1956, The geology of New Hampshire: Pt. 2, Bedrock geology: Concord, N. H., State Plan. Devel. Comm. 203 p.

Boucot, A. J., 1968, Silurian and Devonian of the northern Appalachians, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J.B., eds., Studies of Appalachian geology -- northern and maritime: New York, Interscience Publishers, p. 83-94.

Cady, W. M., 1968, The lateral transition from the miogeosynclinal to the eugeosynclinal zone in northwestern New England and adjacent Quebec, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., eds., Studies of Appalachian geology -- northern and maritime: New York, Interscience Publishers, p. 151-161.

Green, J. C., 1959, Problems in northern New England stratigraphy [abs.]: Minnesota Univ. Center Continuation Study, Inst. Lake Superior Geology, 5th Ann. Mtg., Apr. 13-14, 1959, p. 24.

________________ 1964, Stratigraphy and structure of the Boundary Mountain anticlinorium in the Errol quadrangle, New Hampshire-Maine: Geol. Soc. America Spec. Paper 77, 78 p.

________________ 1968, Geology of the Connecticut Lakes-Parmachenee area, New Hampshire-Maine: Geol. Soc. America Bull., v. 79, no. 11, p. 1601-1638.

Guidotti, C. V., 1965, Geology of the Bryant Pond quadrangle, Maine: Maine Geol. Survey Quad. Mapping Series no. 3, 116 p.

Hall, B. A., 1969, Pre-Middle Ordovician unconformity in northern New England and Quebec, in Kay, Marshall, ed., North Atlantic--Geology and continental drift: Am. Assoc. Petroleum Geologists Mem. 12, p. 467-476.

Harwood, D. S., 1966, Geology of the Cupsuptic quadrangle, Maine: U. S. Geol. Survey open-file report, 259 p.

________________ 1969, The Second Lake anticline--a major structure on the northwest limb of the Boundary Mountain anticlinorium, northern New Hampshire, west-central Maine, and adjacent Quebec: U. S. Geol. Survey Prof. Paper 650-D, p. D106-D115.

Harwood, D. S., and Berry, W. B. N., 1967, Fossiliferous lower Paleozoic rocks in the Cupsuptic quadrangle, west-central Maine: U. S. Geol. Survey Prof. Paper 575-D, p. D16-D23.

Hatch, N. L., Jr., 1963, Geology of the Dixville quadrangle, New Hampshire: New Hampshire Dept. Resources and Econ. Devel. Bull. 1, 81 p.

Moench, R. H., 1969, The Quimby and Greenvale Cove Formations in western Maine: U. S. Geol. Survey Bull. 1274-L, p. L1-L17.

_____________ 1970, Premetamorphic down-to-basin faulting, folding, and tectonic dewatering, Rangeley area, western Maine: Geol. Soc. America Bull., v. 81, no. 5, p. 1463-1496.

Neuman, R. B., 1968, Paleogeographic implications of Ordovician shelly fossils in the Magog belt of the northern Appalachian region, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., eds., Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 35-48.

Osberg, P. H., Moench, R. H., and Warner, Jeffrey, 1968, Stratigraphy of the Merrimack synclinorium in west-central Maine, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., eds., Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 241-253.

Pavlides, Louis, Boucot, A. J., and Skidmore, W. B., 1968, Stratigraphic evidence for the Taconic orogeny in the northern Appalachians, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., eds., Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 61-82.

Rogers, H. D., 1838, Second annual report on the geological exploration of the State of Pennsylvania: Harrisburg, Pa., 93 p.

U. S. Geological Survey, 1961, Geological Survey Research 1961: U. S. Geol. Survey Prof. Paper 424-A, 194 p.

_______________________ 1965, Geological Survey Research 1965: U. S. Geol. Survey Prof. Paper 525-A, 376 p.

Zen, E-an, 1967, Time and space relationships of the Taconic allochthon and autochthon: Geol. Soc. America Spec. Paper 97, 107 p.

_________ 1968, Nature of the Ordovician orogeny in the Taconic area, in Zen, E-an, White, W. S., Hadley, J. B., and Thompson, J. B., eds., Studies of Appalachian geology--northern and maritime: New York, Interscience Publishers, p. 129-140.

Road log.

The route of this trip is in the Rangeley, Oquossoc, and Cupsuptic 15-min. quadrangles, but stops will be made only in the Cupsuptic quadrangle. Assemble at 8:00 a.m. at the junction of Routes 16 and 4 in the center of the village of Rangeley.

Mileage

0.0 Drive west on Routes 16 and 4 toward Oquossoc. Stratigraphic units crossed along this part of the route are discussed in the road logs of trips A-1 and A-3 in this volume.

5.6 Enter Oquossoc quadrangle.

6.5 Turn right (northwest) on Route 16 at the northwest end of Rangeley Lake.

8.1 Cross Kennebago River.

9.7 Enter Cupsuptic quadrangle.

11.7 Turn right (north) from Route 16 at Cupsuptic Nursery.

13.9 Cross Cupsuptic River at the Cupsuptic Sporting Camps. Turn left on the west side of Cupsuptic River and proceed southwest toward the Moochers Home.

14.0 Park cars in the fields west of Moochers Home and continue on foot along an old road along the west side of Cupsuptic River.

14.8 STOP 1. NO HAMMERING PLEASE. We are on the 1740-foot knob on the northwest-trending ridge between the Cupsuptic River and Cold Brook. Outcrops on the top of the knob are rusty-weathering black phyllite, the lowest lithologic unit in the upper member of the Albee Formation. Overlying the black phyllite is green, gray, and purplish-gray phyllite containing less than 10 percent of interbedded "pinstripe" arenaceous rocks that make up most of the upper member of the Albee. Below the black phyllite is greenish-gray and green phyllite containing more than 10 percent of "pinstripe" feldspathic quartzite typical of the bulk of the Albee Formation.

The minor folds in the "pinstripe" arenaceous beds below the black phyllite are of particular interest because they clearly show two widely divergent directions of folding. Axial surfaces of the early folds are variable in trend but generally strike west northwest and dip about 50° S. Axes of these early folds plunge from 10° to 20°E.

to as much as 30°W. and describe a helical pattern across the axial trace of the late folds. Axial surfaces of the late folds are more consistent in strike, being about N. 20° E., but the axes vary in plunge from 10° to 25° NE. to as much as 50° SE. Individual beds traced around this exposure give a three-dimensional picture of dome, basin, and saddle-shaped fold segments typical of superimposed, nearly orthogonal, steeply dipping fold systems.

The regional slaty cleavage is parallel to the axial surface of the late folds. There is no recognizable regional cleavage parallel to the axial surface of the early folds, but a northwest-trending cleavage is faintly visible in the cores of some early folds in this outcrop.

Retrace marked trail back to the cars and retrace the route along the Cupsuptic Tote Road to Route 16.

17.9 Turn left (east) on Route 16.

18.2 Turn left (north) from Route 16 onto private road to north end of Kennebago Lake.

20.1 Cross Kennebago River.

20.3 Turn left (north) onto abandoned Maine Central Railroad grade and proceed north toward Kennebago Lake.

25.5 STOP 2. The gorge of the Kennebago River below the dam at Kennebago Lake is cut in red slate and interbedded "pin-stripe" feldspathic quartzite of the lower member of the Albee Formation. Note that, except for the color caused by microscopic plates of hematite, the red slate and feldspathic quartzite are identical to the green slate and arenaceous rocks of the typical Albee seen at STOP 1. Downstream near the lower power dam are thin discontinuous layers of greenstone in the red slate and a few highly folded beds of relatively clean quartzite about 2 feet thick.

The minor folds on the east side of the river about 100 feet below the upper dam are of particular interest to this trip. Nearly all the minor folds in this large exposure are isoclinal and have near vertical axial surfaces that strike N. 50° E. and axes that plunge 50° to 60° SE. Several hogback minor folds have northeast-trending axial surfaces and axes that plunge from 10° NE. to as much as 50°SW. over a distance of 1 to 5 feet. There is a pronounced cleavage parallel to the northeast-trending axial surfaces of the minor folds but no other cleavage or deformed lineations in this area.

The hogback folds could be the result of differential flowage in the "a" tectonic direction or the result of superimposed northeast-trending late folds on early northwest-trending folds. The fracture pattern near the hinge lines of some of the smaller hogback folds and their diapiric appearance favors differential flowage, but in view of the superimposed folds seen at STOP 1 and the steep plunges of surrounding folds, the alternative interpretation cannot be entirely ruled out.

25.5 Turn vehicles around and retrace the route south along the railroad grade.

29.9 Cross Kennebago River and park vehicles in the abandoned gravel pit about 100 yards west of the bridge. Walk north along the west bank of the Kennebago River to the first outcrops.

30.0 STOP 3. The first outcrop on the west side of the river is white to light-tannish-gray quartz-pebble conglomerate with a light-yellowish-green chloritic matrix. This rock is identical to the quartz-pebble conglomerate member of the Rangeley Formation found near Johns Pond, and thus it is considered to be late Llandovery in age. No fossils have been found at this locality, however, and this isolated sliver of conglomerate cannot be traced into the dated rocks.

A few tens of feet upstream from the first outcrops, the contact between the quartz-pebble conglomerate and green phyllite of the upper part of the Albee Formation is exposed near the river bank. The contact strikes N. 75° E. into the river and is vertical. Bedding is not visible in the Albee, so the existence of an unconformity cannot be unequivocally established.

The quartz-pebble conglomerate is lithologically similar to much of the Clough Quartzite (Billings, 1937, 1956) in New Hampshire. To the east the conglomerate rests on the Albee Formation north of Johns Pond and above polymict conglomerate of the Rangeley Formation southwest of Johns Pond.

Return to the vehicles and retrace the route south along the private road to Route 16.

31.9 Turn left (east) on Route 16 at the Forest Service station.

33.5 Enter the Oquossoc quadrangle.

35.0 Cross the Kennebago River.

35.2 Turn left (north) from Route 16 onto dirt road that follows the abandoned Maine Central Railroad grade.

36.5 Enter the Cupsuptic quadrangle. Low hogback ridge east of the railroad grade is greenstone assigned to the mafic metavolcanic member of the Dixville Formation.

37.3 Turn right (east) from railroad grade onto logging road to Kamankeag Pond.

38.0 Park vehicles on the side of the logging road and prepare for a hike of about 2 miles to the low hills northwest of Kamankeag Pond.

40.2 STOP 4a. Soon after leaving the cars we crossed from the mafic metavolcanic member of the Dixville into the black slate member. We are presently at fossil locality O-1 (fig. 1). The slate here contains a collection of graptolites indicative of Berry's zone 12 of the late Middle Ordovician (fossil list given in Harwood and Berry, 1967). The rock is predominantly black slate that contains scattered beds of feldspathic quartzite as much as 2 inches thick and a few scattered lenses of metavolcanic rocks and some stringers of black chert. The beds are folded but in general strike about N. 40° E. and dip 40° to 80° SE.

Along the slopes to the northwest are scattered exposures of calcareous lithic graywacke that contains beds of granule to cobble conglomerate and thin beds and wispy lenticles of black slate. Clasts in the conglomerate bed are gray quartzite, vein quartz, and chert that are contained in a matrix of graywacke or mixed graywacke and black slate. The calcareous lithic graywacke is a lenticular unit within the black slate member. The upper contact is a gradational zone of alternating graywacke and black slate beds a few inches thick. In the one outcrop where the lower contact is exposed, there is a sharp but conformable change upward from black slate to pebble and cobble conglomerate.

There are abundant scattered fragments of various fossils in the calcareous lithic graywacke. None of the fragments could be identified specifically, but R. B. Neuman of the U. S. Geological Survey tentatively identified one brachiopod fragment as *Bilobia* or *Diambonia*, either of which is compatible with the Middle Ordovician age established by the graptolites.

The layer of calcareous lithic graywacke and included conglomerate is interpreted to be an interformational flysch deposit because of the shallow-water nature of the fossil fragments and the chaotic nature of the included wispy lenticles of black slate.

STOP 4b. We have walked northwestward from the graptolite locality and are now at the first outcrop of cobble-to-boulder polymict conglomerate of the Rangeley Formation. The contact between the basal Silurian rocks and the underlying Middle Ordovician rocks is not exposed but probably is in the low swale southeast of this outcrop.

Bedding is not well developed in the basal part of the polymict conglomerate, which is a jumble of cobbles and boulders composed of vein quartz, gray quartzite, chert, black slate, green slate, minor greenstone, and white-weathering fine-grained felsic metavolcanic rocks and some pebbles of medium-grained gray granite. Several boulders of coarsely crystalline gray limestone that contain recrystallized Halysites and horn corals were collected at this locality. The matrix is feldspathic quartzite, and thus it is quite different from the graywacke-black slate matrix of the Middle Ordovician conglomerate seen at STOP 4a.

Above the basal 50 feet the polymict conglomerate becomes distinctly bedded; beds of cobbles alternate with beds of coarse-grained feldspathic quartzite, both ranging in thickness from 1 to 5 feet. The clasts tend to decrease in size, and the amount of stable quartz, quartzite, and chert relative to unstable slate fragments increase. Granitic cobbles tend to be somewhat more abundant in the middle and upper parts of the polymict conglomerate in this region.

Black slate fragments are distinctly cleaved, but the orientation of their cleavage is essentially parallel to the cleavage in the Silurian rocks, indicating that the cleavage was imposed on the slate chips after they were incorporated in the conglomerate. This finding is consistent with the absence of a pervasive cleavage in the pre-Silurian rocks that can be related to Taconic deformation. The black slate fragments are not deformed around adjacent clasts, however, this implies that their protolith was indurated probably to mudstone or argillite before Silurian time.

Several brachiopod fragments identified as Eoplectodonta or Plectodonta by A. J. Boucot and a few corals were found at locality S-2 (fig. 1). Neither these fossils nor the recrystallized corals from the limestone

boulders positively date the conglomerate but all are compatible with late Llandovery (C_4 - C_5) age fossils found by E. L. Boudette. This is the last STOP of the trip. Follow the marked trail back to Kamankeag Pond and retrace the logging road to the cars.